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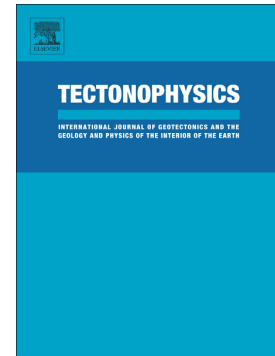
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UPLIFT HISTORY OF A TRANSFORM CONTINENTAL MARGIN REVEALED BY THE STRATIGRAPHIC RECORD: THE CASE OF THE AGULHAS TRANSFORM MARGIN ALONG THE SOUTHERN AFRICAN PLATEAU

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Abstract

The south and southeast coast of southern Africa (from 28°S to 33°S) forms a high-elevated transform passive margin bounded to the east by the Agulhas-Falkland Fracture Zone (AFFZ). We analysed the stratigraphic record of the Outeniqua and Durban (Thekwini) Basins, located on the African side of the AFFZ, to determine the evolution of these margins from the rifting stage to present-day. The goal was to reconstruct the strike-slip evolution of the Agulhas Margin and the uplift of the inland high-elevation South African Plateau. The Agulhas transform passive margin results from four successive stages:

- (1) Rifting stage, from Late Triassic to Early Cretaceous (~200?-134 Ma), punctuated by three successive rifting episodes related to the Gondwana breakup;
- (2) Wrench stage (134-131 Ma), evidenced by strike- and dip-slip deformations increasing toward the AFFZ;
- (3) Active transform margin stage (131-92 Ma), during which the Falkland/Malvinas Plateau drifts away along the AFFZ, with an uplift of the northeastern part of the Outeniqua Basin progressively migrating toward the west;
- (4) Thermal subsidence stage (92-0 Ma), marked by a major change in the configuration of the margin (onset of the shelf-break passive margin morphology).

Two main periods of uplift were documented during the thermal subsidence stage of the Agulhas Margin: (1) a 92 Ma short-lived margin-scale uplift, followed by a second one at 76 Ma located along the Outeniqua Basin and; (2) a long-lasting uplift from 40 to 15 Ma limited to the Durban (Thekwini) Basin. This suggests that the South African Plateau is an old Upper Cretaceous relief (90-70 Ma) reactivated during Late Eocene to Early Miocene times (40-15 Ma).

Keywords: transform passive margin, sequence stratigraphy, South African Plateau uplift

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1. Introduction

Passive margins can be classified according to their tectonic setting (divergent or transform - e.g. Bally and Snelson, 1980), their offshore depositional profile and their onshore topography (occurrence of elevated plateaus or not - e.g. Kooi and Beaumont, 1994; Japsen et al., 2012; Green et al., 2013).

Transform margins can be elevated margins (the south and southeast margins of southern Africa, the margins of Madagascar, the eastern margin of India etc.) but most of them are low-lying (see Mercier De Lépinay et al., 2016 for a review). In some cases, conjugated transform margins can have plateaus on one side and a flat topography on the other side. The best examples are the margins bounded by the Agulhas-Falkland Fracture Zone (AFFZ), with a high-elevated topography to the north (the South African Plateau), and a flat topography below sea level to the south (the Falkland/Malvinas Plateau).

Passive margin tectonic and stratigraphic models were developed in the 1970s (see Bond and Kominz (1988) for a review) as a consequence of plate tectonic models. These models predicted a planation of the relief inherited from the rift shoulders and a progressive overlap of these former reliefs by continental and marine deposits (e.g. Gilchrist and Summerfield, 1991; Ollier and Pain, 1997). During the same period, and based on various approaches, some researchers began to understand that large-scale up to continental-scale uplifts were possible. However, no link between these uplifts and passive margins was made at that time. Progress in the interpretation of apatite fission tracks in the 1990s (e.g. Gallagher et al., 1998; Gunnell et al., 2003) led to a major change in our way of thinking with the characterization of periods of rapid cooling (and erosion) long after the rifting phase. Two main types of models were proposed mainly in

Australia and southern Africa: (1) an uplift due to the isostatic rebound at the time of the escarpment retreat from the initial scarp of the rift shoulders (e.g. Karner and Watts, 1982; Braun and Beaumont, 1989; Weissel and Karner, 1989; Gilchrist and Summerfield, 1991) and (2) a post-rift uplift due to various mechanisms at the mantle or lithosphere scale (e.g. Burke and Gunnell, 2008; Colli et al., 2014; Braun et al., 2014). Even if the first type of model can exist, recent geological data favoured the second type of model along different margins of the world.

We focused here on one of the highest elevated passive margins in the world (more than 3000 m in elevation), the Agulhas transform margin, located on the eastern side of the South African Plateau. The objective of this study is to reconstruct the timing of both the strike-slip evolution of the margin and the uplift of the South African Plateau by analysing the stratigraphic record of the Outeniqua and Durban (Thekwini) Basins. This study is based on a sequence stratigraphic analysis using industrial seismic lines and wells.

2. Geological setting

2.1. Kinematic evolution of the African and South American Plates along the Agulhas Falkland Fracture Zone (AFFZ) and associated magmatism

The Agulhas Margin is conjugate to the Falkland/Malvinas Margin. Together they form the longest transform continental margin in the world (around 1300 km). The first oceanic accretion (Goodlad et al., 1982), marking the onset of the transform movement along the AFFZ, started in the northern Natal Valley between magnetic anomalies M10 and M11 (i.e. 133.58-135.32 Ma - Gradstein et al. 2012) (Figure 1.B). According to the kinematic model of Reeves (2018), the final separation of the continental

Falkland/Malvinas Margin from the southern tip of Africa occurred during the lower Albian (110 Ma). The mid-oceanic spreading centre clears the southern tip of Africa at about 90 Ma shortly before magnetic anomaly C34.

The Agulhas transform margin has experienced periods of intraplate magmatism since the Early Jurassic (Figure 1.B), (1) 184-176 Ma - the Karoo Magmatic Province (LIP) (Duncan et al., 1997; Jourdan et al., 2008; Svensen et al., 2012), (2) 145-133 Ma - the Bumbeni-Movene/Mpilo volcanic system (Allsopp et al., 1984; Saggerson and Bristow, 1983), (3) 135-105 Ma - the Chilwa Alkaline Province (Eby et al., 1995; Woolley, 1991), (4) 105-80 Ma - the Upper Cretaceous Kimberlite Province related to the 'African Superplume' (Jelsma et al., 2004).

2.2. The South African Plateau uplift: present-day knowledge

2.2.1. Morphology

The South African Plateau (or Kalahari Plateau) is a large-scale plateau (elevation of 1000 to 1600 m in South Africa), extending from South Africa to southern Congo, bordered on its seaward side by marginal bulges defining an intracontinental basin, the Kalahari Basin. Marginal bulges are bounded from the coastal lowlands by one or more escarpments. The elevation of the bulges can reach up to 3500 m in the Drakensberg area (Figure 1).

Two major drainage systems cross the plateau: (1) the Limpopo River flows eastward into the Indian Ocean (Figure 1.A); (2) the Orange River flows westward into the Atlantic Ocean, draining a large portion of the plateau interior. Many authors argue for significant drainage reorganizations since the breakup of Gondwana (e.g. Dingle and Hendey, 1984; de Wit, 1999; Moore and Larkin, 2001).

2.2.2. Age of the uplift

The South African Plateau forms a very long wavelength topographic anomaly (> 1000 km) that is postulated to extend offshore across the southeastern Atlantic Ocean. This anomaly is regarded as the expression of the underlying mantle associated with the African Superplume (e.g. Burke and Wilson, 1972; Nyblade and Robinson, 1994; Gurnis et al., 2000; Nyblade and Sleep, 2003). However, the age of the uplift of the plateau is still debated.

2.2.2.1. Thermochronometry approach

Several low temperature thermochronological (mainly apatite fission track analysis, AFTA) studies were performed along the Indian Ocean side of the South African Plateau (Brown et al., 2002; Green et al., 2015; Tinker et al., 2008a; Wildman et al., 2015), characterizing four regional cooling events interpreted as exhumation episodes: 145-130, 110-80, 80-75, and 30-20 Ma. These denudation pulses were correlated with sedimentary flux increases over the margins of southern Africa (Tinker et al., 2008b; Rouby et al., 2009; Guillocheau et al., 2012; Braun et al., 2014; Said et al., 2015).

2.2.2.2. Geomorphological approach

Several geomorphological studies using the analysis of onshore stepped planation surfaces were performed in southern Africa since the pioneering works of L.C. King. One of their objectives was to constrain the uplift history of the South African Plateau (with poor arguments regarding the dating of the erosional surfaces).

King (1948, 1949, 1982) defined three main stepped pediplains of Jurassic ('Gondwana Surface'), Late Cretaceous to Eocene ('African Surface') and Late Cenozoic ('Post-African Surface') age. Each pediplain was supposed to be related to pulses of tectonic uplift.

Partridge and Maud (1987) modified King's original model and recognized three major planation events during: the Early Cretaceous to Miocene ('African Surface'), Miocene ('post-African I Surface'), and Pliocene ('post-African II Surface'). For these authors, the 'African Surface' is a remnant of relief initiated during the continental breakup of Gondwana, and uplifted during the Early Miocene and Late Pliocene times. In their Africa synthesis, Burke and Gunnell (2008) suggested that the Africa topography is inherited from (1) the Gondwana breakup (planation of the 'African Surface') and (2) a still ongoing uplift initiated during the uppermost Eocene (~ 30 Ma). Based on an analysis of river profiles, Roberts and White (2010) support a Late Cenozoic (< 30 Ma) uplift of the South African Plateau.

2.3. Sedimentary record of the margin: previous studies

2.3.1. Outeniqua Basin

The Outeniqua Basin (McMillan et al., 1997) is structured by a set of four half-grabens with an E-W to NNW-SSE orientation (Figure 3.B) with associated depocenters that are called from, east to west, the Algoa, Gamtoos, Pletmos and Bredasdorp Basins (collectively known as the Outeniqua Basin). They are bounded by normal faults (i.e. the St. Croix, Port-Elisabeth, Gamtoos and Plettenberg Faults) and are separated by prominent basement arches (respectively Recife, St. Francis and Infanta). Their southern extension is called the South Outeniqua Basin delimited to the south by the Diaz Marginal Ridge (Figure 3.A.B).

The rifting of the margin (Figure 1.C) started at least during Oxfordian times up to the Late Valanginian (i.e. 160 to 134 Ma). The rift infilling results from two retrogradational sequences mainly composed of continental clastic deposits passing laterally to lacustrine

and/or shallow marine environments (see Dingle et al., 1983 and McMillan et al., 1997 for a review).

From the Late Valanginian to the top Aptian (Bate and Malan, 1992; McMillan et al., 1997; McMillan, 2003; Paton et al., 2004), the structural and sedimentary evolution of the margin was controlled by the drift of the Falkland/Malvinas Plateau along the AFFZ (see § 2.1) which induced differential uplift and subsidence. Deep canyons were scoured in the E-NE part of the margin, whereas thick deltaic sediments (Sunday River Fm) were deposited to the west (Bate and Malan, 1992). After the Falkland/Malvinas Plateau cleared the tip of Africa during the Late Cenomanian (see § 2.1), fault activity rapidly decreased and a passive margin was formed (McMillan, 2003; McMillan et al., 1997).

2.3.2. Transkei Swell

The Transkei Swell extends from the NE part of the Outeniqua Basin to the southern part of the Durban (Thekwini) Basin along the AFFZ. Therefore, it has a narrow continental shelf and steep slope. The deep offshore extension beyond the AFFZ corresponds to the Transkei Basin and the Natal Valley (Figure 3).

Along the coast, two half-grabens (Mbotyi and Mngazana grabens) filled by Late Valanginian sediments (Karpeta, 1987; McLachlan et al., 1976) display an arcuate shape in relation with the AFFZ motion (Figure 3.A).

Onshore, Late Cretaceous and Cenozoic sediments (Niemi et al., 2000) onlap the basement toward the continent at a time of three preserved flooding events (Broad et al., 2012, 2007): (1) Campanian (shallow marine deposits of the Mzamba Fm (Klinger and Kennedy, 1980) and Igoda Fm (Klinger and Lock, 1978), (2) Middle Eocene (carbonate platform deposits of the Bathurst Fm (le Roux, 1990), (3) Plio-Pleistocene

(shallow marine deposits of the Alexandria Fm) in the Algoa Basin (e.g. Roberts et al., 2007).

2.3.3. Durban (Thekwini) Basin

The Durban (Thekwini) Basin filled an asymmetric half-graben (Broad et al., 2012), located in the northern termination of the AFFZ (Ben-Avraham et al., 1997) (Figures 2.B and 3.C). The main border fault is located landward of a prominent basement high, interpreted as a block of Jurassic basalts (Karoo Province), rifted away from the Lebombo Mountains by a NW-SE directed extension during the initiation of the AFFZ motion (Ben-Avraham et al., 1997). The spatial distribution of the graben and onshore faults (Ben-Avraham et al., 1997; Von Veh and Andersen, 1990; Watkeys and Sokoutis, 1998) displays an arcuate shape related to the dextral shear stress along the AFFZ (Figure 3.C). The syn-rift period is at least Late Hauterivian in age (syn-rift II according Broad et al., 2012), whereas the age of their syn-rift I period is unknown.

The post-rift infilling of the margin is poorly constrained. However, Hicks and Green (2016) defined six post-rift seismic units bounded by sequence boundaries. They show a progressive migration of the accommodation area toward the continent in a ramp margin setting during the Cretaceous and Paleocene periods followed by the propagation of a progradational sedimentary shelf during the end of the Paleogene and Neogene.

Based on high-resolution seismic profiles Martin and Flemming (1988) have shown that almost the entire sea floor of the continental shelf of the Durban Basin corresponds to an angular unconformity. Strata from the Upper Cretaceous (Cawthra et al., 2012) to Oligocene (du Toit and Leith, 1974) are tilted toward the basin and truncated by the sea

floor. A similar angular unconformity outcrops in the Zululand (Kwazulu) Basin (see next section).

The present-day morphology of the basin is dominated by the Tugela Delta and Cone (Figure 4.A) which correspond to a large constructional deltaic system that extends across the shelf to a water depth of 3000 m.

2.3.4. The Kwazulu-Maputaland Margin

The Kwazulu-Maputaland Margin extends from Richards Bay (South-Africa) to the Maputo area (Mozambique). The basement is composed of the Karoo volcanic series from the Lebombo Mountains (see § 2.1 for the timing) which forms the western limit of the margin (Figures 2.A and 3.A).

Two poorly known grabens are located beneath that margin in the Lake St Lucia area (McMillan, 2003) and Palmeira area (Salman and Abdula, 1995). They belong to the graben province of the Limpopo Plain which is thought to be Late Jurassic to lowermost Cretaceous in age (Salman and Abdula, 1995). The northeast-trending Bumbeni Ridge, extending from the Lebombo Mountains into the Zululand (Kwazulu) Basin, divides the post-rift sedimentary succession into two depocenters (Figure 3.A).

The Cretaceous sedimentary infill (Zululand Group) (Kennedy and Klinger, 1975; McMillan, 2003; Shone, 2006) is seaward tilted and crops out in the western area of the basin (Figure 3.A). Three main units bounded by hiatuses and discontinuities are defined.

1. The Makatini Fm (= Maputo Fm in the South Mozambique Basin) of Barremian to Early Aptian age, is comprised of alternations of siltstones and shelly nodular limestones

(shallow marine) with fine-grained sandstones (deltas) at the base passing laterally to conglomeratic fan deltas.

2. The Mzinene Fm (= Lower Domo Shale Fm in the South Mozambique Basin) is bounded at its base by a major hiatus including Late Aptian and Early Albian times (Aptian-Albian unconformity). The facies are similar to the underlying formation with shallow marine alternations of siltstones and shelly nodular carbonates with more sandy levels during Middle Albian times.

3. The Santa Lucia Fm (= Domo Sands, Upper Domo Shales and Lower Grudja Fms in the South Mozambique Basin), Coniacian to Maastrichtian in age, is bounded at its base by an angular unconformity with a hiatus from Early Cenomanian to Early Coniacian times (Kennedy and Klinger, 1971). In the distal borehole section, this unconformity is time equivalent to Early Turonian coarse-grained sandstones to conglomerate shore deposits (McMillan, 2003). These sandstones are correlated with (i) the braided river deposits of the Boane Fm (Förster, 1975a) located in north-western Maputaland and (ii) the deltaic (delta front) Domo Sand Fm (Salman and Abdula, 1995) beneath the Limpopo Plain.

Close to the present-day shoreline, Middle Eocene shallow marine bioclastic to oncoidic limestones (Salamanca Fm) (Förster, 1975b) are locally preserved in southern Maputaland. They represent the southern limit of a large carbonate platform (Cheringoma Fm) located beneath the Limpopo Plain (e.g. Förster, 1975a; Salman and Abdula, 1995).

The overlying Neogene succession is bounded by a spectacular angular unconformity ranging from at least the Maastrichtian (and probably Middle Eocene, the geometrical setting of the Salamanca Fm is unknown) to the Early/Middle Miocene (Frankel, 1966). The overlying unconformable deposits are thin remnants of shallow marine bioclastic

limestones (Frankel, 1972, 1968, 1966) ranging from Early/Middle Miocene to Late Miocene.

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3. Material and methods

3.1. Data

Three types of data were combined: reflection seismic profiles, industrial wells (well logs, cuttings, biostratigraphic report), and outcrops. The seismic reflection database consists of an extensive set of offshore industrial 2D seismic reflection profiles (around 32 000 km), made available by the company TOTAL, shot between the 1980's and 2000's. Onshore field data (mainly in the Kwazulu-Maputaland Margin between Richards Bay and Maputo) and thirty-three exploration wells have been used to constrain our seismic interpretation in terms of lithology, age and depositional environments.

3.2. Seismic Stratigraphy

The stratigraphic analysis of the seismic profiles is based on the principles of seismic stratigraphy using two complementary methods: (1) the classical approach originally defined by Mitchum et al. (1977) based on the characterization of seismic facies and the stratal termination patterns, i.e. onlap, toplap, downlap and erosional truncation, and (2) the migration of the shoreline (offlap-break) through time (Helland-Hansen and Gjølberg, 1994; Helland-Hansen and Hampson, 2009). The objective is to define depositional sequences, the stacking of a genetically related regional package called system tracts (Vail et al., 1977) bounded by key surfaces, resulting from changes in the balance between accommodation space (the space created by lithosphere deformation and eustasy - Jervey, 1988) and sedimentary flux (siliciclastic supply from the continent and in situ sediment production, e.g. carbonates). These sequences record a seaward

and landward migration of the shoreline through time, i.e. a progradation (regression) followed by a retrogradation (transgression).

We use the terms formalized by Catuneanu et al. (2009) for both the key bounding surfaces and stratal units (= system tracts). Three main surfaces are defined: (1) sequence boundary (SB) at the beginning of accommodation space removal passing into the continental domain to the unconformity, (2) maximum regressive surface (MRS) at the end of a regression (progradation) and (3) maximum flooding surface (MFS) at the end of a transgression (retrogradation). Four main stratal units are recognized, three (HNR, FR, LNR) during the regression (progradation) and one (T) during the transgression (retrogradation).

- The highstand normal regressive deposits (HNR = highstand system tract or HST of Posamentier et al. (1988)) are the stratal unit bounded between the MFS and the unconformity during the onset of the regression.
- The forced regressive deposits (FR = falling stage system tract or FSST of Plint and Nummedal (2000)) correspond to a basinward shift of the sedimentation contemporaneous with an aerial erosion onshore, the subaerial unconformity (SU).
- The lowstand normal regressive deposits (LNR = lowstand system tract or LST according to the emended definition of Catuneanu et al. (2009)) are the stratal unit bounded between the unconformity and the MRS during the end of the regression.
- The transgressive deposits (T = transgressive system tract or TST of Posamentier et al. (1988)) is bounded by the MRS at the base and the MFS at the top and correspond to the transgression (retrogradation).

- Gravity deposits mainly occurred at time of the forced regressive deposits (FR) as coarse-grained lobes (= basin-floor fan or BFF of Posamentier et al. (1988)) overlapped by slope fan during the LNR and early T deposition on the shelf.

Formally depositional sequences are defined between two unconformities (= sequences boundary) whereas genetic sequences or stratigraphic cycles are defined between two MFS.

Depositional sequences are hierarchized according to their duration (Vail et al., 1991). Here, we focused mainly on second-order sequences (duration of several tens of million years) dominated by tectonic during the sequence boundary formation (Jacquin and de Graciansky, 1998; Vail et al., 1991), nevertheless, some third-order sequences (duration of several millions years) of eustatic origin (Boulila et al., 2011; Strasser et al., 2000) are characterized.

3.3. Age model

The stratigraphic key surfaces defined above (MRS, MFS and SB) have been dated using biostratigraphy.

The biostratigraphic dating is based on the characterization of planktonic foraminifers and calcareous nannofossils biozones of four wells (JC-A1, JC-D1, Zululand 1 (ZU1), ZOE-C) and outcrops from Need's Camp (East London) and Bela Vista (South Mozambique) (see Figure 3.A for location). Lists of microfossils from the wells were re-evaluated by biostratigraphic experts from TOTAL and a new biostratigraphic study of the outcrops was performed by C. Bourdillon (Eradata Service Company). They provide new information on the age and depositional environments of the sedimentary basins on the east coast of South Africa (Transkei Swell, Durban (Thekwini) and Zululand (Kwazulu Basins)). This biostratigraphy revaluation has not been carried out for the Outeniqua

Basin. For this basin, we used the chronostratigraphic framework established by Brown Jr. et al. (1995) and McMillan (2003).

To increase the time resolution, known eustatic events were characterized within the time intervals provided by the biostratigraphy. Oxygen isotopes curves were used as a proxy of the eustasy (Friedrich et al., 2011; Zachos et al., 2001).

3.4. Isochore (thickness in TWT) maps

Interpolated isochrone and isochore maps (in milliseconds two-way travel time or ms TWT) of six second-order unconformities were calculated. They result from a kriging interpolation of our 2D seismic dataset integrated into the 'Sismage' software suite (TOTAL's proprietary software).

3.5. Uplift characterization

An uplift of a passive margin can affect the whole margin or parts of it (e.g. only the inland erosional domain). The stratigraphic record of an uplift has three distinct signatures: (1) a tilting of the margin sediments truncated by an angular unconformity, (2) a major relative sea level fall and (3) an increase in the siliciclastic sediment supply.

The tilt of a margin results from an uplift of the inner domain (coastal plain and upstream catchments) and an increase in the subsidence in the outer domain. The uplifted area is eroded and overlapped by subtabular sediments truncating underlying tilted sediments (angular unconformity).

An uplift of the margin can also lead to a major relative sea-level fall and a sharp downward shift of the shoreline defined as a forced regression (Helland-Hansen and Hampson, 2009; Helland-Hansen and Martinsen, 1996; Posamentier et al., 1992)

corresponding to the falling stage system tract (FSST) of Posamentier and Allen (1999) or to the forced regressive deposits (FR) of Catuneanu et al. (2009). To discriminate tectonic forcing from the eustatic component in a downward migration of the shoreline, the amplitude of the relative sea level fall must be higher than the maximum rate of sea level fall (around 100 m of sea level fall over a 1 myr period according Miller et al. (2005) and Bessin et al. (2017)).

An increase in the siliciclastic sediment supply can record an uplift as well as a major climate (precipitation) change or a drainage reorganization (e.g. drainage capture). This criterion is not enough to discriminate an uplift.

4. Durban (Thekwini) Basin fill evolution

The Durban (Thekwini) Basin is located over a half-graben (Figure 4) with the main border fault located landward of a prominent basement high (outer shelf basement high). In the inner domain the acoustic basement is composed of continental crust and folded sediments from the Cape Fold Belt (Cape Supergroup and Dwyka Tillite) as indicated by the JC-A1 well (Figure 5). Basinward, the nature of the basement is not drilled. However the seismic facies are similar, suggesting that the acoustic basement could be stretched continental crust.

4.1. Seismo-stratigraphic framework (second-order sequences - several 10 myr): distribution, architecture, sedimentology and structural deformation

Eight second-order sequences (duration of several 10 myr) were defined. Their main characteristics including ages are summarized on Table 1 (supplementary material) and Figures 4 and 5. Six thickness maps (in ms TWT) are presented in Figures 3.A and 6:

Total thickness; Berriasian-Late Valanginian; Late Valanginian-Early Turonian; Early Turonian-top Maastrichtian; top Maastrichtian-Late Eocene; Late Eocene-Present-day.

4.1.1. Second-order Sequence S1: ?-134 Ma (Berriasian to Late Valanginian)

Sequence S1 is the first sequence infilling the half-graben described previously. Seismic geometries show evidence of growth-strata indicating syn-rift deposition (Figure 4). The upper part of S1 was drilled in the JC-B1 well (Figure 5). It consists of fine to coarse-grained sandstones intruded by dolerite intrusions. The age and depositional environment of the sequence are unknown (barren of fossils).

A similar graben exists in the southern part of the Kwazulu-Maputaland Margin north of Richards Bay. It was drilled by the Zululand 1 (ZU 1) well (see Figure 3.A for location) 160 km away from the JC-B1 well. The graben infill consists of 790 m of grey silty claystones grading up to siltstones and fine sandstones with basalt pebbles and some lignite. This succession is interpreted as continental deposits for the first 120 m (absence of dinoflagellates in the palynological record) grading into mixed shallow marine and continental environments. Biostratigraphy data (revaluated here) gives an age ranging from Berriasian to Hauterivian.

On the S1 isochore map (Figure 6) the main depocenter is located on the hangingwall of the graben described earlier with up to 500 ms TWT (around 1000 m) of accumulated sediments in agreement with the structural map of Ben-Avraham et al. (1997).

4.1.2. Second-order Sequence S2: 134-113 Ma (Late Valanginian to top Aptian)

Sequence S2 has a wedge-shaped geometry (Figure 4) in which the thickness reaches up to 600 ms TWT (around 1150 m) into the central part of the graben. Internal reflections

show sub-parallel geometries onlapping the fault plane and hanging wall dip-slope. The upper part of sequence S2 is drilled in the JC-B1 well (Figure 5). It consists of Aptian sand-rich coarsening-upward sediments (from fine sandstones to conglomerates) interpreted as mouth bar deposits that pass upward to siltstones. Dolerites intrusions cut through these deposits.

The Late Valanginian sequence boundary is interpreted as the breakup unconformity (BUU) synchronous to the seafloor spreading initiation in the northern Natal Valley during the magnetic anomaly M10 (~ 134 Ma) (Goodlad et al., 1982).

4.1.3. Second-order Sequence S3: 113-92 Ma (top Aptian to Early Turonian)

The base unconformity (SB3) of sequence S3 corresponds to a hiatus that spans the lower Albian in the JC-B1 well (Figure 5). The sequence is subdivided into two shorter duration sequences (third-order?).

4.1.3.1. Sequence S3a: 113-101 Ma (top Aptian to Late Albian)

Two system tracts are defined (Figure 4): a basin-floor fan time-equivalent of the FSST (FR) and an HST (HNR). The basin-floor fan corresponds to a sedimentary wedge with sub-parallel reflections, probably turbidite deposits, onlapping the degraded fault scarps both landward and toward the ocean. It records the last infill stage of a remnant rift topography. The highstand system tract (HST) is a progradational-aggradational delta as indicated by the JC-B1 well (delta-front facies, Figure 5).

4.1.3.2. Sequence S3b: 101-92 Ma (Late Albian to Early Turonian)

Sequence S3b records a major shift of the depocenters from the middle to the outer part of the margin at time of the FSST (FR). The highstand system tract (HST) is preserved on the inner margin as a prograding delta. This is both a major landward shift of the depocenters and the first sediment deposition of the upstream basement (mfs3b). The delta is made up (well JC-B1) of clayey siltstones grading upward to fine to medium sandstones interpreted as prodelta slope deposits (Figure 5).

The isochore map (Figure 6) indicates that sequences S2 and S3 filled a topography inherited from rifting. The mapping of the offlap-break (shoreline) for the prograding wedge of HST (S3b) shows the occurrence of two deltas.

4.1.4. Second-order Sequence S4: 91-73 Ma (Early Turonian to Late Campanian)

The sequence boundary SB4 is an erosional unconformity bounding, on the inner margin, tilted truncated clinoforms and the overlying sediments of sequence S4 (Figure 4). Along the outer margin, SB4 is overlapped by a basin-floor fans (BFF) shifted far beyond the outer shelf basement high (Figure 4). The next lowstand system tract (LST or LNR) forms an aggradational to progradational delta (up to 940 m in the JC-B1 well) (Figure 5) characterized by well-developed prodelta slope deposits (clayey siltstones in the JC-A1 and JC-B1 wells). The highstand system tract (HST) is restricted to the inner margin and is composed of two stacked progradational deltaic wedges (lower order sequences) (Figure 4).

4.1.5. Second-order Sequence S5: 73-66 Ma (Late Campanian to top Maastrichtian)

The SB5 unconformity is overlapped basinward by backstepping basin-floor fans (BFF) deposits (Figure 4). The lowstand system tract (LST) is located down-dip than the previous one of sequence S4 but shows similar facies. The highstand system tract (HST) is characterized by condensed levels on the inner margin passing upstream to low angle deltaic clinoforms.

The isochore map of sequences S4 and S5 (Figure 6) shows the individualization of two depocenters on both sides of the previous syn-rift subsiding domain. To the north, the mapping of the offlap-break (shoreline) of HST S5 confirms the presence of a deltaic system prograding toward the SE. To the south, the second depocenter overlaps the previously formed oceanic crust (see the structural framework in Figure 3.C).

4.1.6. Second-order Sequence S6: 66-40 Ma (top Maastrichtian to Early Bartonian)

The stratal pattern is very similar to those of sequences S3 to S5 (Figure 4). Unconformity SB6 is overlapped by backstepping basin-floor fans (BFF) and overlapped by prograding deltaic systems with well-developed prodeltaic slope deposits (silty claystones with very-fine-grained sandstones on JC-A1 and JC-B1) of the LST (up to 835 m).

The main difference with sequences S3 to S5 lies in the highstand system tract (HST) formed by a progradational-aggradational delta with evidence of gravitational deformation (large slumps) (Figure 4). The maximum flooding surface (mfs6, base of this HST of Lutetian age) is a major turn-around at the margin-scale. It records the end of an overall aggradation-retrogradation since the breakup unconformity around 134

Ma (Late Valanginian) and the beginning of an overall progradation of the margin until today.

The isochore map of sequence S6 (Figure 6) shows a change in the configuration of the depocenters with a southward migration of the deposition area. Sediments were probably supplied by a main river located south of Durban and a smaller one along the coast between Durban and Richards Bay (proto-Tugela) with a strong influence of waves as indicated by the narrow elongated distribution of the sediments along the offlap-break.

4.1.7. Second-order Sequence S7: 40-26 Ma (Early Bartonian to top Oligocene)

Sequence S7 is subdivided into two upper order sequences (third-order?) bounded by two sequence boundaries (SB7 and SB7b). As in sequence S6, significant erosional truncations are evident on the upstream part of the basin below the sea floor (Figure 4). The timing and implications of this will be discussed in the next section.

4.1.7.1. Sequence S7a: 40-33 (Early Bartonian to intra Rupelian)

Unconformity SB7 is highly erosional, truncating both the LST and HST of the previous sequence S6. This unconformity is overlain in the outer margin by backstepping basin-floor fans (BFF) (Figure 4). The overlapping LST (215 m thick in the JC-A1 well) forms a deltaic progradational wedge made up of an alternation of silty claystones and calcareous sandstones with abundant shell fragments organized in coarsening-upward cycles (Figure 5). Basinward, the delta front facies pass to delta slope environments (JC-B1 well, Figure 5) and along the outer margin large mounded forms (Figure 4), which are interpreted as contourite deposits (elongated mounded drifts of Stow et al. (2002)).

The next highstand system tract (HST) records a major change in the sedimentary evolution with the first occurrence of carbonate platform deposits (prograding distally steepened carbonate platform) made up (JC-A1 well, 130 m thick) of an alternation of calcareous claystones and sandy and shelly limestones (Figure 5).

4.1.7.2. Sequence S7b: 33-26 (intra Rupelian to Chattian)

This sequence is restricted to the inner margin. No basin-floor fans occurred. The progradational wedge of the lowstand system tract (LST) consists (JC-B1 well, 150 m thick) of an alternation of calcareous claystones and sandy limestones deposited along a carbonate slope (Figure 5). Elongated mounded drift occurred along the outer margin (Figure 4).

The next highstand system tract (HST) forms a progradational-aggradational carbonate ramp. It consists (JC-A1 well, 230 m thick) of massive bioclastic limestones of Oligocene age deposited in a high-energy inner ramp setting (Figure 5).

4.1.8. Second-order Sequence S8: 26-0 Ma (Chattian to Present)

Sequence S8 is subdivided into three upper order sequences bounded by three unconformities (SB8, SB8b and SB8c).

4.1.8.1. Sequence S8a: 26-23 Ma (Late Chattian)

Sequence S8a is onlapping landward onto unconformity SB8 and is then restricted to the middle to outer margin. Sequence SB8a does not exist in the JC-A1 well where an uppermost Chattian hiatus is documented for unconformity S8a (Figure 5). This might be the age of the sequence.

Both the lowstand system tract (LST) and highstand system tract (HST) are carbonate ramps made up (JC-B1 well) of sandy limestones (LST) or alternations of sandy limestones with fine-grained sandstones (HST). The slope of the topsets of the progradational wedge of the HST is not as inclined as the one of the previous sequence S7b (Figure 4).

4.1.8.2. Sequence S8b: 23-20 Ma (top Chattian to top Burdigalian)

Sequence S8b facies are more siliciclastic than the underlying sequence S7 and S8a. The depositional environment seems to be a mixed system with little deltas interbedded with bioclastic carbonates as indicated by the 100 m-thick lower Miocene sandy limestones of the JC-A1 well (du Toit and Leith 1974) or the fine- to medium-grained sandstones with shell fragments of the JC-B1 well (Figure 5).

Capping an onlapping lowstand system tract (LST), the highstand system tract (HST) forms a nice prograding deltaic system.

4.1.8.3. Sequence S8c: 20-0 Ma (top Burdigalian to Present)

Unconformity SB8c forms a major unconformity (major downward migration of the shoreline) with no basin-floor fan preserved at the toe of the deltaic slope.

This sequence is poorly constrained both in age (no cuttings available in the JC-B1 well), and in term of seismic imaging (importance of the multiple and quality of the seismic processing). McMillan (2003) described thin sediments in the JC-A1 well unconformably overlying the Early Miocene (Burdigalian) sandy limestones of S8b. These thin sediments are correlated with the onshore unconformable bioclastic sandy

limestones of the Uloa Formation dated Langhian in age (Frankel, 1968, 1966). This unconformity occurred around the Burdigalian-Langhian boundary.

Geometrically this sequence is located down-dip from the last offlap-break (shoreline) of HST S8b as a large lowstand wedge. The first downward progradational deltaic unit onlapping the unconformity forms a pure progradation with no evidence of forced regression. This unit is then intermediate between a FSST (FR) and a LST. The next highstand system tract (HST), poorly imaged, seems to be an aggradational to progradational wedge.

4.2. Sedimentary and deformation evolution of the Durban (Thekwini) Margin

4.2.1. 145? to 134? Ma (Berriasian? to Late Valanginian?): Syn-Rift

The rift period is characterized by a narrow (width of 15 km), low subsiding (maximum thickness around 1000 m with a mean sedimentation rate of 80 m/myr), and asymmetric half-graben. The main border faults dip landward and define a prominent basement high. The age of the syn-rift sedimentation is poorly constrained and ranges from Berriasian to Late Valanginian according to new ages available from the Zululand 1 well (see above). The breakup unconformity is thought to be contemporaneous with the first oceanic accretion recorded between magnetic anomalies M10 and M11 (133.58-135.32 Ma, Late Valanginian - Gradstein et al., 2012) in the northern Natal Valley (Goodlad et al., 1982, Martin et al., 1982) immediately southward of the outer basement high described earlier (Figure 3.C).

4.2.2. 134? to 113 Ma (Late Valanginian? to top Aptian): planation of the Rift topography

The stratal pattern is characteristic of a flexural regime (sag basin type) with sub-parallel strata onlapping both the hangingwall dip-slope and the degraded fault scarps (Figure 4). The thickness map (Figure 6) suggests that this period represents the passive infill phase of remnant rift topography in a post-tectonic setting, corresponding to the immediate post-rift system tract (as defined by Prosser (1993)).

The sedimentary system is characterized by narrow deltas in shallow marine conditions infilling the inherited rift topography.

4.2.3. 113 to 92 Ma (top Aptian to Early Turonian): initiation of the passive margin topography

The Aptian-Albian unconformity (SB3 ~ 113 Ma) corresponds to a hiatus extending from Late Aptian to base Albian. Based on seismic observations, the hiatus is conformable and does not show any truncation features meaning deformations with a wavelength higher than the basin. This hiatus has a regional extension. It was recognized in the Zululand (Kwazulu)-Maputaland Basin (Kennedy and Klinger, 1975; Förster, 1975a).

Above, there is a sedimentary wedge onlapping toward the continent. This unit is interpreted as a late post-rift system tract (*sensu* Prosser, 1993) corresponding to the last infilling of the rift topography (Figure 4).

The planation of the tilted blocks was probably completed during the maximum flooding event at the Albian-Cenomanian transition (~ 105 Ma) when the Durban Graben was totally filled and formed a ramp margin (Figure 4) in a shallow water setting as indicated by the height of the clinoforms (around 200 m) (Figure 7).

4.2.4. 92 Ma (Early Turonian): first short-lasting uplift of the margin

The Early Turonian unconformity (SB4 ~ 92 Ma) is a regionally extensive erosional discontinuity that is interpreted as a record of a significant regional uplift event (Figure 7):

- Offshore, this unconformity initiates a major change in the topography of the margin with a deepening of the outer part of the Durban (Thekwini) Basin (lower slope to abyssal depositional environment, benthic foraminifers of JC-B1) and the deposition of the first deep sea fan sediments. Toward the inner margin, the Cenomanian deltaic topsets are truncated.
- Onshore, this unconformity could be related to a sharp erosional event recorded in the Zululand (Kwazulu)-Maputaland Basin with the deposition of Turonian? coarse-grained fluvial sandstones of the Boane Fm (Förster, 1975a; Kennedy and Klinger, 1971; McMillan, 2003).

The biostratigraphic reevaluation of the ZOE-C well (see Figure 3 for location) in the Zululand (Kwazulu) Basin gives an age ranging from 91 to 93 Ma with no evidence of a hiatus. This suggests a quite short duration for this Early Turonian unconformity.

4.2.5. 92 to 43 Ma (Early Turonian to Lutetian): high topography margin

Thermal post-rift subsidence prevailed during the Late Cretaceous and Early Paleogene with a progressive landward migration of the shoreline until the Lutetian marine maximum flooding event (~ 43 Ma). The sequences (S4, S5, S6) display the same stratigraphic pattern (basin-floor fan, lowstand system tract disconnected from the shelf and highstand system tract restricted to the shelf). The stacked deltas of the inner

margin are parallel with no evidences of truncations indicating the absence of any uplift during this period (Figure 7).

However, the physiography of the margin shows considerable strike variations (Figure 6):

- To the north, at the southern termination of the Zululand (Kwazulu) Basin, a deltaic system progrades over a broad continental shelf (35 km in width) and the slope is gently dipping.
- In the central area (location of the seismic line in Figure 4) the margin topography is more pronounced with a narrow continental shelf (25 km in width) bounded by a non-depositional slope. The reconstruction of the margin depositional profile (Figure 7) indicates a slope height around 1500 m. At the time of the base level fall, sediments bypass the narrow continental shelf to feed the backstepping turbiditic systems of the basin-floor fans (BFF) into the deep sea basin.
- To the south, along the northern part of the Agulhas-Falkland Fracture Zone (AFFZ), the margin is narrow with a low accommodation potential on the shelf. The sediments bypass the narrow shelf (15 km width) toward the deep sea basin, feeding the onlapping basin-floor fans (BFF).

These strike variations in the margin physiography are easily explained by the inheritance of the opening stage of the Natal Valley with three structural domains (Figure 3.C): (1) a low extended continental crust in the northern area, (2) a graben system in the central area and (3) an oceanic crust bounded by the AFFZ in the southern area.

In the absence of basin uplift during this period, the dominant factors for the SB5 (intra Late Campanian) and SB6 (uppermost Maastrichtian) unconformities may be eustatic as

indicated by the major sharp global cooling events characterized by Friedrich et al. (2011) during these periods.

4.2.6. 43 to 34 Ma (Lutetian to Late Priabonian): delta growth

The major Lutetian marine maximum flooding event (basin-scale first-order mfs) of the margin (mfs6 ~ 44 Ma) is contemporaneous with the growth of carbonate platforms outcropping southward along the Transkei Swell (Bathurst Fm) (le Roux, 1990), northward in the northern part of the Zululand (Kwazulu)-Maputaland Basin (Salamanca Fm) (Förster, 1975b) and in the Limpopo Basin (Cheringoma Fm) (Flores, 1973; Salman and Abdula, 1995). This Lutetian age was confirmed by our biostratigraphic reevaluation of the famous Need's Camp outcrop near East London along the Transkei Swell and the Bela Vista quarry in the Maputaland Basin (see Figure 3.A for location).

In the Durban (Thekwini) Basin, the Lutetian maximum flooding marks the onset of a large delta in the central and southern part of the studied area (Figure 6), which probably recorded a proto-Tugela River. The Early Bartonian unconformity (SB7 ~ 40-41 Ma) is coeval with an increase in the siliciclastic sediment input and the formation of a turbiditic system (basin-floor fan).

The change in the stratal pattern of the margin, at the first order (one or two hundreds of millions years), going from retrogradational to progradational (sensu Vail et al., 1977, 1991; Duval et al., 1998) or from underfilled to overfilled (sensu Henriksen et al., 2011) during the Lutetian, is coeval with the onset of the Zambezi and Limpopo Delta (Salman and Abdula, 1995).

In the deepest part of the basin, the occurrence of mounded drifts due to contourite records an increase in the deep sea oceanic circulation (Antarctic Bottomwater) as evidenced by Uenzelmann-Neben et al. (2007).

4.2.7. 34 to 23 Ma (Late Priabonian to top Chattian): development of Carbonate platforms

The Oligocene time interval is a singular period with the second occurrence of a carbonate platform. These carbonate sediments are not known onshore. They are contemporaneous with the large siliciclastic deltas of the Limpopo and Zambezi Rivers.

4.2.8. 23 to 0 Ma (top Chattian to Present): second uplift of the margin

The S7 unconformity is a major seaward tilting of sediments unconformably capped by subtabular Middle Miocene deposits. Along the shelf of the Durban (Thekwini) Basin tilted truncated seaward-dipping Cenozoic strata directly outcrop on the sea floor (firstly recognized by du Toit and Leith (1974) and mapped by Martin and Flemming (1988)). Onshore the Zululand (Kwazulu) Basin (Frankel, 1972, 1968, 1966; McMillan, 2003) and the Maputaland (South Mozambique) Basin (Frankel, 1972), tilted seaward-dipping sediments ranging from Barremian (Makatini Fm) (Kennedy and Klinger, 1975; McMillan, 2003; Shone, 2006) to Lutetian (Salamanca Fm) (Förster, 1975b) are truncated by tabular Middle Miocene carbonate deposits (Uloa Fm) (Frankel, 1972, 1968, 1966). The biostratigraphic dating of these former carbonates suggests that they could be the onshore equivalent of the maximum flooding surface mfs8 ~15 Ma characterized in the Durban (Thekwini) Basin.

This angular unconformity is a smooth marine planation surface (wave-cut platform) extending throughout the Durban (Thekwini), Zululand (Kwazulu,) and South

Mozambique Basins. This unconformity fossilized a major regional flexure (doming) of the margin (Figure 7). Compared to the major previous uplift (SB4, Early Turonian), the duration of this uplift is longer as indicated by the progressive change in the slope of the topsets during Early Miocene times. This stratal pattern indicates an acceleration of the flexuration seaward of the hinge zone and an uplift landward of the hinge zone. It ends before the eustatic Middle Miocene marine flooding when shallow carbonate deposits of this age capped the planated unconformity.

The present-day Tugela Delta and Cone is initiated after this major Middle Miocene marine flooding and is recorded up to the present-day as a large progradational wedge. Unfortunately, the poor quality of the seismic data in the shallowest part of the basin does not allow the characterization of the major climatic/eustatic changes in the base Pleistocene and the Late-Middle Pleistocene Transition.

5. Outeniqua Margin fill evolution

5.1. Nature of the acoustic basement and main structures

The regional seismic section (Figure 8) crosses over the Pletmos graben which is bordered by two basement highs: the Infanta Arch landward and the Diaz Marginal Ridge basinward (Figure 3.B). The basement of the inner and middle parts of the margin (Figure 8) is composed of shales and quartzites of the Cape Supergroup (Dingle et al., 1983). The Diaz Marginal Ridge was not drilled, however seismic facies suggest that it could be a continental crust intruded by numerous magmatic sills

Extensional structures

The four NW-SE trending half-grabens (Figure 3.B) were filled by the syn-tectonic sediments of sequence S1 and later tilted toward the S-SW (Figure 8).

Strike-Slip structures

Toward the AFFZ fracture zone, the rift faults display an arcuate shape (Figures 3.B and 10) compatible with a right-lateral deformation. Southward, these faults are connected with E-W *en-échelon* strike- and dip-slip faults, oblique to the axis of the Diaz Marginal Ridge. These strike-slip structures were filled by the syn-tectonic sediments of sequence U2a (Figure 8).

5.2. Seismo-stratigraphic framework (second-order sequences - several 10 myr): distribution, architecture, sedimentology, and structural deformation

Seven second-order sequences (duration of several 10 myr) were defined. The unconformities were named using the SOEKOR (today PetroSA) terminology (see Brown Jr. et al., 1995). Their main characteristics including ages, are summarized in Table 2 (supplementary material) and Figures 8 and 9. Six thickness maps (in ms TWT) are represented in Figures 3.A and 10: Total thickness; Late Valanginian-Aptian; Aptian-Early Turonian; Early Turonian-Campanian; Campanian-top Maastrichtian; top Maastrichtian-Present-day.

5.2.1. Second-order sequence S1: ?-134 Ma (Middle Jurassic to Late Valanginian)

Sequence S1 shows evidence of growth strata indicating a syn-rift deposition. A detailed description of the syn-rift stratigraphy of the Outeniqua Basin is beyond the scope of this paper (see Dingle et al., 1983 and Paton and Underhill, 2004 for a review).

5.2.2. Second-order sequence S2: 134-113 Ma (Late Valanginian to top Aptian)

This interval can be split into two sedimentary sequences bounded by two erosional surfaces: 6At1 and 13At1 unconformities. The first sequence is very condensed or absent over a large part of the margin. The top of the \pm S2 isochore map corresponds to the 13At1 surface. Thus sequence S4, bounded at the top by the 14At1 unconformity, is not completely imaged whereas the \pm S3 isochore map includes the uppermost part of the S2 Sequence.

5.2.2.1. Sequence U2a: 134-131 Ma (Late Valanginian to Late Hauterivian)

The base unconformity 1At1 is of Late Valanginian age (McMillan, 2003) and corresponds to the breakup unconformity (Bate and Malan, 1992; du Toit, 1977; McMillan et al., 1997) that marks a change in the stratal pattern between rift infilling to the transitional to early drift period. The main depocenters of sequence U2a (Figure 10) are located along the syn-rift hangingwall faults (Plettenberg Fault: up to 3200m, the Gamtoos Fault: up to 2500m). Some structures are inverted (anticline folds) during this time interval in the Gamtoos and Bredasdorp Basins as already described by Van Der Merwe and Fouche (1992) and Paton and Underhill (2004).

Along the outer margin (Southern Outeniqua Basins), two E-W trending *en-échélon* depocenters are observed oblique to the axis of the Diaz Marginal Ridge (DMR) (Figure 10, thickness map \pm S2). The sedimentary infilling (Figure 8.B) has a wedge geometry thickening toward the DMR, with sub-parallel reflectors onlapping both fault planes of the DMR and the hangingwall slopes toward the continent. Unfortunately the depositional environment is not constrained. However, McMillan et al. (1997) described deep marine claystones in the inner margin suggesting that the seismic geometries

described above could correspond to deep sea fan deposits. A sedimentary source from the DMR cannot be excluded.

5.2.2.2. Sequence U2b: 131-113 Ma (Late Hauterivian to top Aptian)

The base unconformity 6At1 is of Hauterivian to Early Barremian age (McMillan, 2003). It records both the end of the wrench deformation of the margin and the initiation of a prograding shelf. Unconformity 6At1 is over most of the Outeniqua Basin and is an erosional hiatus (6At1 = 1At1). The Outeniqua Basin subsidence pattern changes from NE to SW (Figure 10).

- To the southwest, along the Bredasdorp and Pletmos Basins, the main depocenters are filled in the inner margin by large progradational deltas (Figure 8) consisting (Ga-W1 well, Figure 9) of coarsening-upward alternations of siltstones to medium-grained sandstones. These deltaic facies are correlated with the Sunday River Fm. Four third-order sequences (bounded by unconformities 6At1, 7At1, 9At1 and 11At1) are characterized with forced regression wedges (FFST or FR) above the 7At1 and 9 At1 unconformities. In map view (Figure 10), the offlap-break of the Sunday River delta, here the transition between shelf and shallow marine/fluviatile sediments (ramp margin setting), indicates the occurrence of two deltaic sediment feeders in the Bredasdorp and Pletmos Basins. The capping retrogradational trend (TST or T) is made up of clayey siltstones on wells Ga-W1 and Ga-S1 (Figure 9). The overlying maximum flooding surface (13At1) is an organic-enriched clayey layer (McMillan, 2003; McMillan et al., 1997).

- To the northeast, along the Gamtoos and Algoa Basins, sediment deposition is low. The degree of erosion of the 6At1-1At1 unconformity increases eastward, east of the Recife Arch, experiencing deeper erosion than the Gamtoos Basin. Two canyons

were scoured: a little one in the Gamtoos Basin close to the Gamtoos Fault and a much larger one (75 km long, 25 km wide, and up to 1 km deep) incised in the Uitenhage Trough (Figure 10). These canyons were filled by shelf and slope deposits from 6At1 to 13At1. The end of the canyon infilling corresponds to a flat erosional unconformity (smooth shallow marine planation surface = wave-cut platform) truncating a substantial area of the syn-rift sequence in the Algoa and Gamtoos Basins. The Canyon incision is related to the uplift of the Uitenhage and Port Elizabeth Trough probably planated by the same wave-cut platform as the one on the top of the canyons.

5.2.3. Second-order sequence S3: 113-92 Ma (top Aptian to Early Turonian)

Sequence S3, bounded by the 14At1 unconformity, started by coarse-grained turbiditic sandstone (Ga-S1 well) deposits of the basin-floor fan (BFF) overlain by a forced regression wedge (FSST) with a major downward shift of the offlap-break, evidencing a major relative sea level fall (Figure 8). The above retrogradational trend (TST or T) is truncated in the middle margin and probably fully eroded in the inner margin (Figure 8). The \pm S3 isochore map shows lateral thickness variations with maximum values in the Bredasdorp Basin (Figure 10). In this area, the thick clastic infill results from the deposition of basin-floor fans and progradational to aggradational sandy deltas both confined along the Bredasdorp Basin axis (NW to SE).

5.2.4. Second-order sequence S4: 92-76 Ma (Early Turonian to Upper Campanian)

The base unconformity (15At1) is erosional with truncations in the inner and middle margin (Figure 8). The overlying stratal pattern displays an overall progradation with

three units bounded by the unconformities 15At1, 15Ct1, 17At1, corresponding to upper order (third?-order) sequences.

The first one is a third-order (?) highstand system tract (HST or HNR). The second one (bounded by unconformity 15Ct1) is a forced regression (FSST or FR) followed by a lowstand system tract (LST or LNR) and a transgressive system tract (TST or T). The last and largest package is a spectacular progradational-aggradational wedge (LST or LNR) made up of several stacked LST of upper order sequences (Figure 8). These progradational units are composed of clayey siltstones (Ga-S1 well) on both the clinofolds and the topsets, meaning that the offlap-break should be interpreted as a shelf break rather than the deltaic shoreline.

Unconformity 15At1 is one of the major stratigraphic discontinuities in the Outeniqua Basin evolution. The isochore map shows that the physiography inherited from the rifted basins (recognizable until the Late Cenomanian) was completely cleared by that time with the progradation of a more linear continental shelf along the margin (Figure 10).

5.2.5. Second-order sequence S5: 76-66 Ma (Upper Campanian to top Maastrichtian)

Unconformity 18A1 is erosional basinward and in the bottomset part of the previous sequence S5. In some areas (Gamtoos and Algoa Basin), the erosion can be very significant with truncations cutting down the syn-rift deposits.

The unconformity records a major relative sea level fall (Figure 10) as indicated by a major downward shift of the offlap-break. The following lowstand wedge (LST or LNR) is built downward of the large silty wedge of sequence S4. The LST is coarsest-grained (Ga-S1 well) and displays a coarsening-up trend (Figure 9). As the upper unit

of sequence S5, several upper order lowstand prograding wedges are stacked here with evidences of gravitational sliding (Figure 8).

Above, the entire margin is overlapped by a large aggradational-retrogradational system (transgressive system tract, TST or T) composed of glauconitic silty claystones (McMillan et al., 1997) capped by a major maximum flooding surface (mfsIG, Figure 8) followed by a highstand system tract (HST) (Figure 8).

The relative sea level fall recorded by the 18At1 unconformity might be correlated with the onshore Igoda incision (see § 2.3.2) infilled by Late Campanian to Early Maastrichtian sediments (Klinger and Lock, 1978). In this scenario, the maximum flooding event mfsIG could be Maastrichtian in age

5.2.6. Second-order sequences S6 and S7: 66 to 0 Ma (top Maastrichtian to Present-day)

In the absence of biostratigraphic well dating, the age of the sediments of sequences S6 to S7 is not known. On the inner and middle margin, two Cenozoic prograding wedges overlapping the 22At1 unconformity and made up of alternations of claystones, limestones, and glauconitic sandstones (McMillan, 2003) are interpreted as a highstand system tract (HST). The two HST are separated by an erosional hiatus that may range from the Oligocene to Early Miocene (Dingle, 1973).

Seaward of the present-day shelf break, strong sea floor erosion by oceanic current is illustrated by the truncations of the 22At1 unconformity. Evidence of truncation in the outer margin during the shaping of the Eocene-Oligocene unconformity indicates that oceanic currents were already active at the time of the Eocene-Oligocene transition (Figure 8). A recent submarine erosional area is clearly recognizable on the S6 and S7 isochore maps (Figure 10). It occurs at a water depth between 1000 and 1700 m. The

erosion seems to be more pronounced toward the southwest with the formation of an erosional concave contouritic channel. The depocenter located in the eastern part of the margin is attributed to contouritic mounded drifts.

5.3. Sedimentary and deformation evolution of the Outeniqua Margin

5.3.1. 160? to 134 Ma (Oxfordian? to Late Valanginian): Syn-Rift

The syn-rift infill of the Outeniqua Basin comprises two sedimentary sequences, not studied here, related to two extensional phases (Dingle et al., 1983; Bate and Malan, 1992; McMillan et al., 1997; Paton and Underhill, 2004).

1. This first cycle (Oxfordian? to Late Kimmeridgian-Early Tithonian - Paton and Underhill, 2004) made of continental sediments with one marine flooding (Early Kimmeridgian), is fault-controlled with wedge shape depocenters thickening toward the hangingwall dip-slope of the Algoa, Gamtoos and Pletmos grabens.

2. The second cycle (Late Kimmeridgian-Early Tithonian to Late Valanginian) is characterized by a generalization of the extension at the margin-scale with the initiation of the Bredasdorp Basin. The sedimentary infill is still continental with conglomeratic alluvial fans along the faults close to sea level as testified by marine floodings.

At the end of the rifting, the margin is structured into four half-grabens (width around 100 km) bounded by normal faults whose footwall formed prominent basement arches.

5.3.2. 134 to 92 Ma (Late Valanginian to Early Turonian): active transform continental margin and first uplift of the margin

The structural setting of the Outeniqua Basin is typical of a continental transform margin (e.g. Basile et al., 1993) with a marginal ridge (Diaz, DMR) and structures (see

above) compatible with a dextral motion of the AFFZ. Two periods with different deformation regimes may be defined:

1. before the South Atlantic breakup (Late Hauterivian, ~ 131 Ma), a wrench period, reactivating inherited half-grabens in anticlinal folds and the former normal faults in dip- and strike-slip motion, and creating E-W *en-échélon* syn-tectonic basins along the northern edge of the DMR;
2. after the South Atlantic breakup (131 Ma), a decrease in the fault activity when the newly created oceanic crust came into contact with the AFFZ. The Outeniqua Margin became an active transform margin.

This second period marks a change in the wavelength of the deformation. It is coeval with an uplift of the Algoa and Gamtoos Basins (NE) with canyon incision and subsidence in the Pletmos and Bredasdorp Basins (SW) with the deposition of thick deltaic wedges (Sunday River Fm). The flattening of the relief created around the Algoa and Gamtoos Basins and the infilling of the surrounding canyons occurs at 124 Ma (Late Barremian, 13At1). Nevertheless, subsidence remained located along the central part of the margin (124-113 Ma, 13At1-14At1) and migrated westward (Bredasdorp Basin, 113-94 Ma, 14At1-15At1) (Figure 10).

5.3.3. 92 Ma (Early Turonian): second uplift of the margin

As mentioned previously, the base Turonian unconformity (15At1 ~ 92 Ma) is a major unconformity with a change in the morphology of the margin. The topography inherited from the rifting and the transform motion of the AFFZ was cleared and a shelf break margin was formed with the height of clinofolds reaching 600 m.

This regionally extensive erosional unconformity formed in response to a significant regional uplift event (Figure 11) is supported by (1) a tilting of the inner margin, (2) an erosion of Cenomanian sediments deposited along the whole margin (biostratigraphic data of McMillan, 1990 and McMillan et al., 1997) and (3) erosional truncations.

The geometry of the first onlap termination of sequence S4 is parallel, suggesting that significant relief was present before marine flooding.

5.3.4. 92 to 76 Ma (Early Turonian to Upper Campanian): 'passive' margin

The successive passive infill of the margin was formed by progradational wedges, progressively filling the relief created during the Early Turonian uplift.

5.3.5. 76 to 72 Ma (Late Campanian): third uplift of the margin

The 18At1 (~ 76 Ma) unconformity records a major relative fall in sea level during Late Campanian times that can be measured, using the height of the downward migration of the offlap-break, at 400-600 m (i.v. 2200 m/s). Even if no tilting is characterized, this amplitude of relative sea level fall cannot be of eustatic origin only (maximum 100 m). This indicates a very long-wavelength uplift of the whole margin and its hinterland, that might be coeval with the incision of valleys in Igoda along the Transkei Swell (Klinger and Lock, 1978).

5.3.6. 72 to 0 Ma (Late Campanian- Cenozoic): starved margin

This time period corresponds to a sharp decrease in the sediment supply all over the margin. No uplift is recorded as indicated by the absence of tilting of the topsets and

forced regression wedges. The Late Cenozoic (more precise dating not available) is characterized by intense reworking by the oceanic currents.

6. Discussion

The main questions in the understanding of elevated passive margins are the timing and causes of the uplift with respect to the rift period and, in the case of a transform margin, the relationships between strike-slip deformations. For these reasons, we successively summarized and discussed the vertical movements of the Agulhas Margin occurring during (1) the rifting stages leading to the AFFZ initiation, (2) the active transform motion of the AFFZ and (3) the passive transform motion of the AFFZ. These kinematic phases of a transform passive margin were originally defined by Mascle and Blarez (1987).

6.1. Successive failed rifts along the Agulhas-Falkland Fracture Zone (AFFZ).

The rifts contemporaneous with the Gondwana breakup of southern Africa are poorly constrained in age as well as seismic imaging. For these reasons, South American conjugate margins (Argentina Margin and Falkland/Malvinas Plateau) must be included in this discussion.

The pre-transform history of the Agulhas-Falkland/Malvinas transform margin is quite complex, punctuated by three successive rifting episodes related to the Gondwana breakup. The extension probably results from a combination of processes induced by the West Gondwana subduction and possibly related mantle plumes (e.g. Dalziel et al., 2013).

6.1.1. Triassic to Early Jurassic continental rifting

The Triassic to Early Jurassic time is a period of widespread extension both in southern South America (e.g. Zerfass et al., 2004) and in Central and Eastern Africa with the Karoo Rifts (e.g. Daly et al., 1992) in response to the post-Gondwanides extension (the Cape Fold Belt part of the external Gondwanides is now dated before 253 Ma, i.e. Permian - Blewett and Phillips, 2016).

Along the studied area, Triassic to Lower Jurassic syn-rift sediments are preserved along onshore outcrops of the San Jorge and Magallanes Basins (e.g. Uliana and Biddle, 1987; Uliana et al., 1989). Offshore, according to Biddle et al. (1986), they may correspond to the deepest infill of the half-grabens seismically characterized in the West Malvinas Basin (Galeazzi, 1998, 1996).

6.1.2. Middle Jurassic-Late Jurassic (i.e. 170-150 Ma) continental rifting

This Middle to Late Jurassic period of extension is more restricted in space to the southern parts of the South American (Franzese et al., 2003) and African Plates with the reactivation of older rift structures (Magallanes and West Malvinas Basins) or with the initiation of several NW-SE trending isolated rifts (West Falkland and North Falkland grabens, Pletmos, Gamtoos, Algoa and Falkland/Malvinas Plateau grabens). This period of extension occurred after the flooding of the basalts from the Karoo Magmatic Province (LIP) around 182-183 Ma in Austral Africa and is contemporaneous with the polygenic large silicic Chon Aike volcanic Province in Patagonia (188 to 153 Ma - Pankhurst et al., 2000). The end of this extensional period is coeval with the first oceanic spreading in the Mozambique Channel that may be Early Oxfordian in age (159 Ma, chron M33n - Leinweber and Jokat, 2012) even if some authors suggest that it might be older (Early Bathonian, 168 Ma, chron M41n - Leinweber et al., 2013). This Middle-Late

Jurassic extension is probably the regional record of the stress change at the origin of the opening of the Weddell Sea and Mozambique Oceanic Basin, from the continental rift to the oceanic spreading (e.g. Jordan et al., 2017; Storey et al., 1996).

To the west (Magallanes and West Malvinas Basins), the rifting is of Middle Jurassic age with a breakup unconformity of Callovian age (Biddle et al., 1986; Galeazzi, 1996, 1998; Bransden et al., 1999). Syn-rift sediments are continental volcanoclastic deposits (part of the Chon Aike Province) of the Tobifera Fm (Biddle et al., 1986; Galeazzi, 1998, 1996). The early rift phase of the North Falkland is not drilled, but a younger age (Oxfordian) is proposed (e.g. Richardson and Underhill, 2002; Lohr et al., 2015).

In the central part of the studied area (Figure 12), within the Pletmos, Gamtoos, Algoa, and Falkland/Malvinas Plateau grabens complex, the syn-rift infill was mainly continental clastic (Biddle et al., 1996; Dingle et al., 1983). Several marine floodings were recorded in these basins: Oxfordian (Maurice Ewing Bank - Thompson, 1977), Kimmeridgian (Gamtoos Basin - McMillan et al., 1997) and onshore in the Pletmos (Knysna Fm - Dingle and Klinger, 1972) and Algoa Basins (Colchester Fm - Dingle et al., 1983).

Grabens assumed to be of the same age were described northward (1) in the Mozambique Basin along the Limpopo coastal plain (N-S grabens filled by undated continental red deposits - Flores, 1964, 1973; Salman and Abdula, 1995 and (2) in the Lower Zambezi graben (rhyolites dated at 166 ± 10 Ma interstratified within continental sandstones of the Lupata Fm - Flores, 1964). New dating is required to confirm the occurrence of Late Jurassic grabens in this area.

6.1.3. Late Jurassic-Early Cretaceous (i.e. 150-131 Ma) continental rifting

This third phase of rifting was initiated during the Late Jurassic with the onset of (1) a N-S purely extensional rift along the proto-South Atlantic divergent margin and (2) NW-SE isolated grabens along the proto-AFFZ (Figure 12).

The N-S rift along the proto-South Atlantic reactivated and/or cut across the earlier North Falkland graben (Brandsen et al., 1999; Lawrence et al., 1999; Lohr and Underhill, 2015; Richardson and Underhill, 2002). The syn-rift infill was characterized by clastic continental deposits and igneous material. Only the uppermost portion of the syn-rift interval has been drilled along the Atlantic Margin of South Africa, providing a Late Hauterivian age (Jungslager, 1999). Biostratigraphic data suggest a Tithonian to Valanginian age in the North Falkland graben (Richardson and Underhill, 2002).

In the Outeniqua area, NW-SE earlier grabens were reactivated together with the onset of the Bredasdorp Basin. The sedimentation is still clastic-dominated, with the deposition of continental to shallow marine sediments of the Kirkwood and Infanta Fms, whereas open marine conditions (anoxic black shales) prevailed in the Falkland/Malvinas Basin (Dingle et al., 1983).

To the east, the rift phase was marked by the onset of isolated half-grabens presently oriented NW-SE along the proto-AFFZ (i.e. Mngazana and Mbotiy grabens) and the proto-Northern Natal Valley divergent margin (i.e. Durban and Zululand grabens). The syn-rift infill was mainly composed of continental to shallow marine clastic deposits accompanied by igneous material, dated from Berriasian to Late Valanginian (wells ZU1 and ZOE-C, this study).

This phase of rifting is the time equivalent of the Agulhas-Falkland/Malvinas transform margin initiation with (1) the Late Valanginian Northern Natal Valley breakup (i.e. 134 Ma) followed by (2) the South Atlantic breakup (i.e. 131 Ma). This event is

contemporaneous with widespread regional magmatic activity in southeast Africa (Bumbeni-Movene/Mpilo Province, see § 2.1) and along the volcanic South Atlantic divergent margin (seaward dipping reflectors (SDR) and Etendeka Province, see § 2.1).

6.2. Vertical movements occurring during the active transform motion of the AFFZ (134-92 Ma): first Cretaceous uplift of the South African Plateau

6.2.1 The wrench phase, from 134 to 131 Ma

The main discontinuities and stratal pattern changes occur around 134 Ma (Late Valanginian - 1At1) along the AFFZ African Margin with (1) reactivation, in the Outeniqua Basin, of the inherited normal faults (graben inversion and strike-slip fault movement Paton and Underhill, 2004) and formation of wrench basins along the DMZ and (2) initiation, in the Durban (Thekwini) Basin, of a flexural regime in a shallow marine setting after the rift phase (§ 4.2.2).

This early post-rift evolution of the margin (Figure 12) results from the drifting of the Falkland/Malvinas continental plateau along the Agulhas-Falkland Fracture Zone (AFFZ) after the opening of the Natal Valley, from Late Valanginian (first oceanic crust in the natal Valley - Goodlad et al., 1982) to Late Cenomanian (final separation of the Falkland/Malvinas Plateau from the Agulhas Bank - Martin et al., 1982). The Outeniqua Margin is therefore an accommodation zone with respect to the steadier Falkland/Malvinas Plateau. The wrench phase is suggested via (1) strike- and dip-slip deformation increasing toward the fracture zone, (2) reactivation of the inner margin inherited rifts and (3) formation of E-W *en-échélon* syn-tectonic basins, perpendicular to the NW-SE Outeniqua grabens, along the Diaz Marginal Ridge. These vertical movements (onset of the *en-échélon* basins) could be expressed by the tilt of the future marginal ridge toward the continent.

6.2.2. Active transforms margin, from 131 to 92 Ma

The northeastern part of the Outeniqua Basin (Algoa and Gamtoos Basins) is uplifted around the Late Hauterivian (~ 131 Ma) (Figure 12). The uplift progressively migrates toward the west as evidenced by the progressive migration of the accommodation area. This migration probably led to additional uplift in the outer area of the margin along the Diaz Marginal Ridge. The newly created relief in the Algoa and Gamtoos Basins were planated around 124 Ma (Late Barremian).

The Aptian-Albian boundary (SB3/14At1) is a major unconformity with a time gap from Late Aptian to Early Albian along the Durban to Zululand-Maputaland Margins (§ 2.3.4. - Kennedy and Klinger, 1975; McMillan, 2003; Shone, 2006), with no evidence of fault activation or tilting, indicating a deformation with a wavelength higher than the basin and without any connection with the transform history of this margin. This deformation is recorded everywhere in Africa (e.g. Guiraud and Maurin, 1991, 1992; Guiraud, 1998) and probably must be related to a major change in plate stresses (so-called Austrian deformations).

The last evidences of wrench tectonic activity (e.g. flexures on both sides of the outer ridge in the Durban (Thekwini) Basin) are sealed by the Early Turonian unconformity (92 Ma): the African side of the AFFZ is no longer a transform margin.

Both the wrench and transform periods are characterized by the progradation of sandy deltaic systems.

6.2.2.1. Comparison with onshore thermal history studies

As mentioned previously (§ 2.2.2.1.), different apatite fission track analyses (AFTA) (e.g. Brown et al., 2002; Tinker et al., 2008a; Wildman et al., 2015) converged toward two Cretaceous cooling periods: 145-130 and 110-80 Ma. Green et al. (2015) focused on the Outeniqua Margin and suggest the occurrence of two exhumation periods (120-100 Ma and 80-75 Ma) during the post-breakup evolution of the margin.

Our interpretations support the scenario put forth by Green and co-authors which suggests the existence of a sedimentary cover over the present-day onshore Outeniqua region that might be, in our interpretation, sediments deposited at the time of the maximum sea extension during Early Aptian times.

The discrepancy between the age of the uplift of the Algoa and Gamtoos Basins coming from our stratigraphic analysis and a cooling event between 120 and 100 Ma in the western part of the margin can be explained by the migration of the uplift toward the west. Therefore, the difference of uplift ages between the two approaches can be consistent.

6.2.2.2. Deformation mechanisms

The observed structures of the wrench phase of the Outeniqua Margin (1At1 to 6At1) are characteristic of the onset of a transform margin (e.g. Basile et al., 1993; Basile et al., 2015). By this time and just before the continental breakup of the South Atlantic, the relative motion between the African and South American plates increases and the Outeniqua area becomes an accommodation area where vertical and horizontal movements occur. One of the consequences is the uplift of the Diaz Marginal Ridge with the formation of E-W *en-échélon* syn-tectonic basins. The evolution of this type of ridge is a matter of debate (see Basile et al., 2015 for a review). Basile et al. (1993) model

suggests that the contact between the colder extended lithosphere (here, the one of the Outeniqua Margin) and the hot oceanic lithosphere could induce a thermal uplift of the ridge that forced the previous tectonic tilting.

The inversion of the eastern part of the margin and the associated uplift probably reflect the proximity of the Algoa and Gamtoos Basins to the AFFZ when the Maurice Ewing Bank began its traverse. The drift of the Maurice Ewing Bank may also explain the progressive migration of the accommodation zone toward the west of the margin and potentially the reactivation of inherited structures as evidenced onshore (Green et al., 2015).

6.3. Vertical movements occurring during the passive transform motion of the AFFZ (92-0 Ma)

6.3.1. The Late Cretaceous uplifts of the South African Plateau

The Outeniqua and Durban (Thekwini) Basins experience a major short-lived uplift with seaward tilting around 92 Ma (base Turonian) (Figure 13) as suggested by (1) topset tilting and truncation of the inner margins, (2) sharp deepening of the outer margins, and (3) forced regression records of a relative sea level fall higher than 100 m (maximum values for eustasy).

Later, both margins have a different evolution: (1) To the south, along the Outeniqua Basin area, the margin experiences a second uplift during Late Campanian times (around 76 Ma, 18 At1) with an amplitude of 400-600 m. (2) To the north, in the Durban Basin, the margin is retrogradational with two low amplitude relative sea level falls that might be of eustatic origin. No uplift is recorded.

The 92 Ma uplift is a major change in terms of the sedimentary system and submarine topography with a major deepening of the middle to outer margin of the Durban

(Thekwini) Basin and a change from a ramp setting with sandy deltaic systems to a true silt-dominated shelf-break margin controlled by thermal subsidence.

No uplift is recorded from 76 Ma (Late Campanian) to around 40 Ma (Bartonian). This is a quite stable tectonic period with the development of carbonate platforms during Middle Eocene times all around southern Africa (Transkei Swell - Bathurst Fm, Durban Basin, Zululand-Maputaland Margin up to Limpopo Plain - Salamanca/Cheringoma Fm).

6.3.2. The Late Eocene? to Early Miocene (40-20 Ma) uplift of the eastern part of the South African Plateau

This uplift is only recorded in the Durban (Thekwini) Basin with a long-lasting progressive uplift from around 40 Ma (poorly dated) to 23 Ma (uppermost Oligocene to transition to Miocene).

The Late Eocene unconformity records major change in the sedimentary systems with the birth of a new siliciclastic system before the growth of a second carbonate platform during Oligocene times. No tilting is recorded. This means a far-field uplift probably enhanced by the major climatic change operating at the Eocene-Oligocene boundary (an aridification here - Tyson and Partridge, 2000).

The uppermost Oligocene is a major relative sea level fall with sharp tilting of the underlying strata and a progressive tilting during Early Miocene times. This uplift ended during uppermost Early to Middle Miocene times as shown by the tabular sediments of this age unconformably overlying tilted Late Cretaceous sediments (Förster, 1975b; Frankel, 1972). Similar observations were made along the Atlantic Margin of South Africa (Stevenson and McMillan, 2004). This means that the very long-wavelength uplift of the South African Plateau occurs at this time.

The only place where uplift occurs after this period is in the coastal plain of the southern Transkei Swell, between Port Alfred and East London, where Miocene and Pliocene sediments occur at elevations up to 300-400 m (more than the eustasy around 50 to 100m) (Maud and Botha, 2000). This uplift may have induced the modern Tugela delta and cone system.

6.3.2.1. Comparison with onshore thermal histories studies

Apatite fission track analyses (AFTA, see § 2.2.2.1) (Brown et al., 2002; Tinker et al., 2008a; Wildman et al., 2015) suggest two regional cooling periods during this time interval, 100-80 Ma and 80-75 Ma. The first cooling episode was evidenced along the southern and southeastern margins of southern Africa, whereas the second one was documented in the southern and southwestern margins of southern Africa. Our results, two uplifts around 92 and 76 Ma, are in agreement with these data. The Eocene-Early Miocene uplift is not documented by low temperature thermochronology except by Green et al. (2015) who modelled a third cooling event along the Outeniqua Margin. The absence of cooling event in the AFTA record in the other parts of the margin might be explained by an uplift with low denudation due to an aridification of the climate since Oligocene times. This can also explain the preservation of such an old relief in southern Africa.

6.3.2.2. Deformation mechanisms

The driving forces for the Late Cretaceous uplifts of southern Africa remains poorly understood (see Rouby et al. (2009) for a review).

The Early Turonian uplift is limited to the Indian Ocean side margin of southern Africa. There is no evidence of uplift during that period along the Atlantic Margin of southern

Africa; this occurred later during the Late Campanian (Braun et al., 2014). The Atlantic side is characterized during Turonian times by a major increase in the siliciclastic sedimentary flux (Guillocheau et al., 2012). This period of uplift, coeval with a major denudation onshore, overlaps (in age) the formation of two large mafic igneous provinces (LIPs) (i.e. Madagascar LIP and Agulhas LIP) and a southern African-scale pulse of Kimberlite activity. These observations, including the size of the uplifted plateau, suggest a relationship between mantle dynamics and the uplift of the southern and southeastern margins of southern Africa. Braun et al. (2014) proposed a model of migration of the African Plate over the South African Superplume with first, at 92 Ma, a westward tilting toward the Atlantic Ocean (explaining the uplift located on the Indian Margin side and the increase in siliciclastic flux) and second, a progressive uplift to the west ending at 76 Ma along the Atlantic Margin.

Nevertheless we cannot exclude a possible role of the AFFZ. In fact, according to Reeves (2018), the passive transform motion of the AFFZ started shortly before the magnetic anomaly C34, i.e. at about 90 Ma contemporaneous to the period of the uplift.

7. Conclusions

The objective of this study is to reconstruct the timing of both the strike-slip evolution of the Agulhas Margin and the uplift of the inland Southern Africa Plateau, analysing the stratigraphic record of the Outeniqua and Durban (Thekwini) Basins located along the AFZZ. This study is based on a sequence stratigraphic analysis. The main results are as follows.

1. A sequence stratigraphic analysis of two basins based on reevaluated biostratigraphic data for the Durban (Thekwini) Basin. Second-order sequences are

defined, 7 in the Outeniqua Basin and 8 in the Durban (Thekwini) Basin. Some bounding unconformities are of the same age over the whole transform margin:

- Late Aptian Unconformity (~ 113 Ma) - SB3 or 14At1.
- Early Turonian Unconformity (~ 92 Ma) - SB4 or 15At1.
- Top Maastrichtian Unconformity (~ 66 Ma) - SB6 or 22At1.

2. A dated reconstruction of the different steps of the opening of the transform margin from Middle Jurassic to Early Turonian. Four steps are defined:

- Middle/Late Jurassic oblique extension (170-150 Ma) restricted to the Outeniqua Margin and the conjugated Falkland/Malvinas Plateau.
- Late Jurassic-Early Cretaceous oblique extension (150-134 Ma) extending all over the margins of southern Africa.
- Early Cretaceous wrench deformation (134-131 Ma) with strike- and dip-slip deformations increasing toward the AFFZ.
- Lower Cretaceous-early Upper Cretaceous active transform margin (131-92 Ma) with an uplift of the northeastern part of the Outeniqua Basin progressively migrating toward the west (migration of the accommodation area).

3. A characterization of two main periods of uplift after the active transform margin period (since 92 Ma):

- An early Late Cretaceous (92 Ma) short-lived margin-scale uplift followed by a second one during Late Cretaceous (76 Ma) located along the Outeniqua Basin;
- Cenozoic long-lasting uplift from 40 to 15 Ma limited to the Durban (Thekwini) Basin;

4. A reconstruction of the successive sedimentary systems along the margin with a major change in configuration at 92 Ma: onset of a shelf-break passive margin controlled by thermal subsidence.
5. The age of the uplift of the South African Plateau is Late Cretaceous, confirming the results from thermochronological data, with a reactivation on the eastern part (Durban area) from 40 to 15 Ma.

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Figures and tables caption

Figures

Figure 1: Topographic and geological setting of the south and southeast passive margins of southern Africa. A. Different types of passive margins with the age of the first oceanic accretion. B. South and southeast passive margins of southern Africa: geomagnetic anomalies on oceanic crust, plateaus, sedimentary basins, magmatism and kimberlite provinces. The purple area refers to the Upper Cretaceous Kimberlite location in southern Africa (Jelsma et al., 2004), consequence of the 'African Superplume' activity between 100 to 80 Ma.

Figure 2: Geometry of the south and southeast passive margins of southern Africa: A. the South-Mozambique Margin, B. the Durban (Thekwini) Margin, C. the Outeniqua Margin (see Figure 1 for location). The three sections are constructed by combining several seismic profiles converted in depth.

Figure 3: A. Mesozoic and Cenozoic coastal basins along the Agulhas Margin (inspired from Ben-Avraham et al., 1997) . B. Structural map of the Outeniqua Basin (modified from McMillan et al., 1997). C. Structural map of the Durban (Thekwini) Basin (Ben-Avraham et al., 1997; Broad et al., 2007).

Figure 4: Line-drawings of a regional seismic profile across the Durban (Thekwini) Basin, based on 2D seismic reflection sections and well interpretations (see Figure 3 for location). A. Regional section with ages. B. Detailed geometry of the margin sedimentary wedge.

Figure 5: Correlation of two wells of the Durban (Thekwini) Basin based on the seismic stratigraphy interpretation (see Figure 4), with mention of the lithology and depositional environments. The age model is based on a reevaluation of the biostratigraphy data for the JC-A1 and JC-D1 wells.

Figure 6: Isochore maps (thickness ms TWT) of six stratigraphic units (see name and position of surfaces in Table 1 (supplementary material) and seismic section in Figure 4).

Figure 7: Evolution of the topography and depositional systems along a profile from the hinterland to the deepest part of the Durban (Thekwini) Basin.

Figure 8: Line-drawings of a regional seismic profile across the Pletmos and Outeniqua Basins based on 2D seismic reflection lines and well interpretations (see Figure 3 for location). A. Regional section with ages. B. Detailed geometry of the margin sedimentary wedge.

Figure 9: Correlation of two wells of the Pletmos Basin based on the seismic stratigraphy interpretation (see Figure 8), with mention of the lithology and depositional environments. The age model is based on McMillan (2003) and McMillan et al. (1997).

Figure 10: Isochore maps (thickness ms TWT) of six stratigraphic units (see name and position of surfaces in Table 2 (supplementary material) and seismic line in Figure 8).

Figure 11: Evolution of the topography and depositional systems along a profile from the hinterland to the deepest part of Pletmos and South Outeniqua Basins.

Figure 12: Synthetic map of the tectonics events occurring in southern Africa during the Middle/Late Jurassic and Early Cretaceous. Plate tectonic reconstruction based on Martin et al. (1981, 1982).

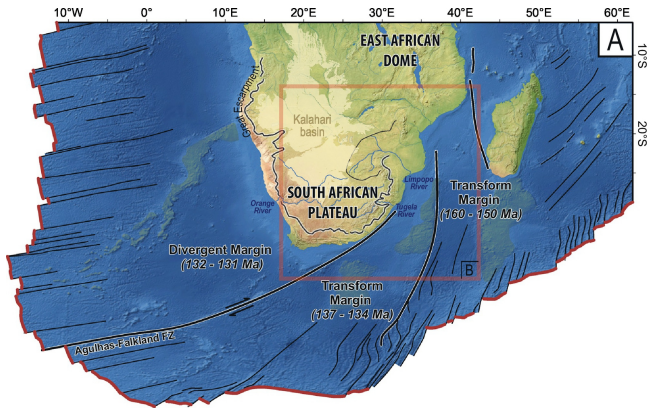
Figure 13: Synthetic map of vertical movements in southern Africa during A. early Late Cretaceous (~ 92 Ma), B. Late Cretaceous (81-66 Ma), C. Paleogene-Neogene boundary (34-26 Ma). The plate tectonic reconstruction on the right side was done using the Gplate software, the plate circuit is modified from Seton et al. (2012). White = continental crust, light blue = oceanic crust, medium blue = Kimberlite provinces, dark blue = large igneous provinces, grey = oceanic plateaus, red lines = uplifted areas.

Figure 14: Synthetic chart of the south and southeast passive margins of southern Africa: magmatism, oceanic accretions, regional scale deformations and stratigraphic record.

RESEARCH HIGHLIGHTS

- The Agulhas transform passive margin results from four successive tectonic stages.
- Rift stage: 200?-134 Ma, wrench stage: 134-131 Ma.
- Active transform margin stage: 131-92 Ma, thermal subsidence stage: 92-0 Ma.
- South African Plateau uplift occurs at 92 Ma (short-lived) and 40-15 Ma (long-lived).
- Uplifts of the plateau are marked by strike, long-wavelength flexures of the margin.

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OFFSHORE CRUSTAL DOMAINS

- Continental
- Basement highs
- Oceanic
- Magnetic anomalies
- Main sedimentary basins (Mesozoic and Cenozoic)
- Rift systems

MAGMATIC PROVINCES

- Neogene
- Late Cretaceous (alkaline Province)
- Lower Cretaceous (Bumbezi-Movene/Mpilo Pr. Chilwa Province)
- Jurassic (Karoo Province)

KIMBERLITES

- Kimberlites distribution
 - Upper Cretaceous Province
- A.I. Alphard Bank
 Ch. Chilwa Province
 D. Drakensberg
 L. Lebombo Mountains
 Z. Zambezi Valley

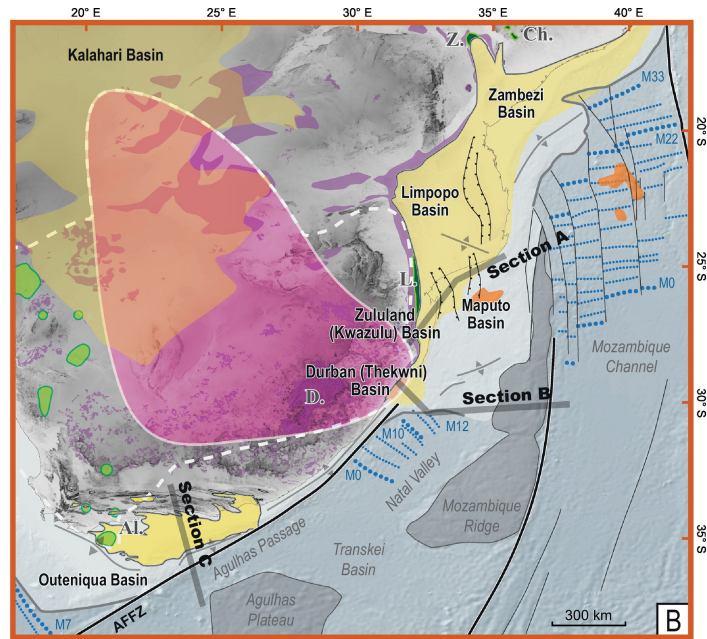


Figure 1

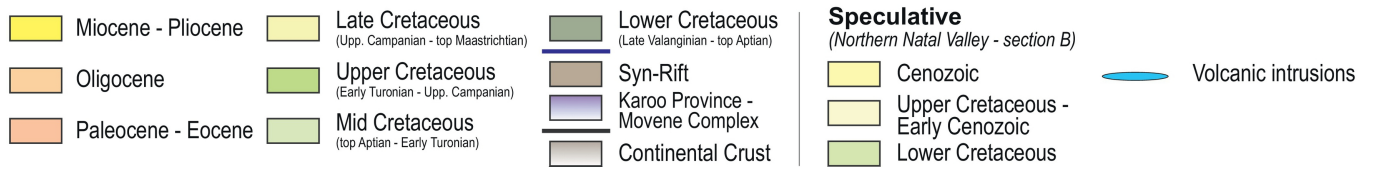
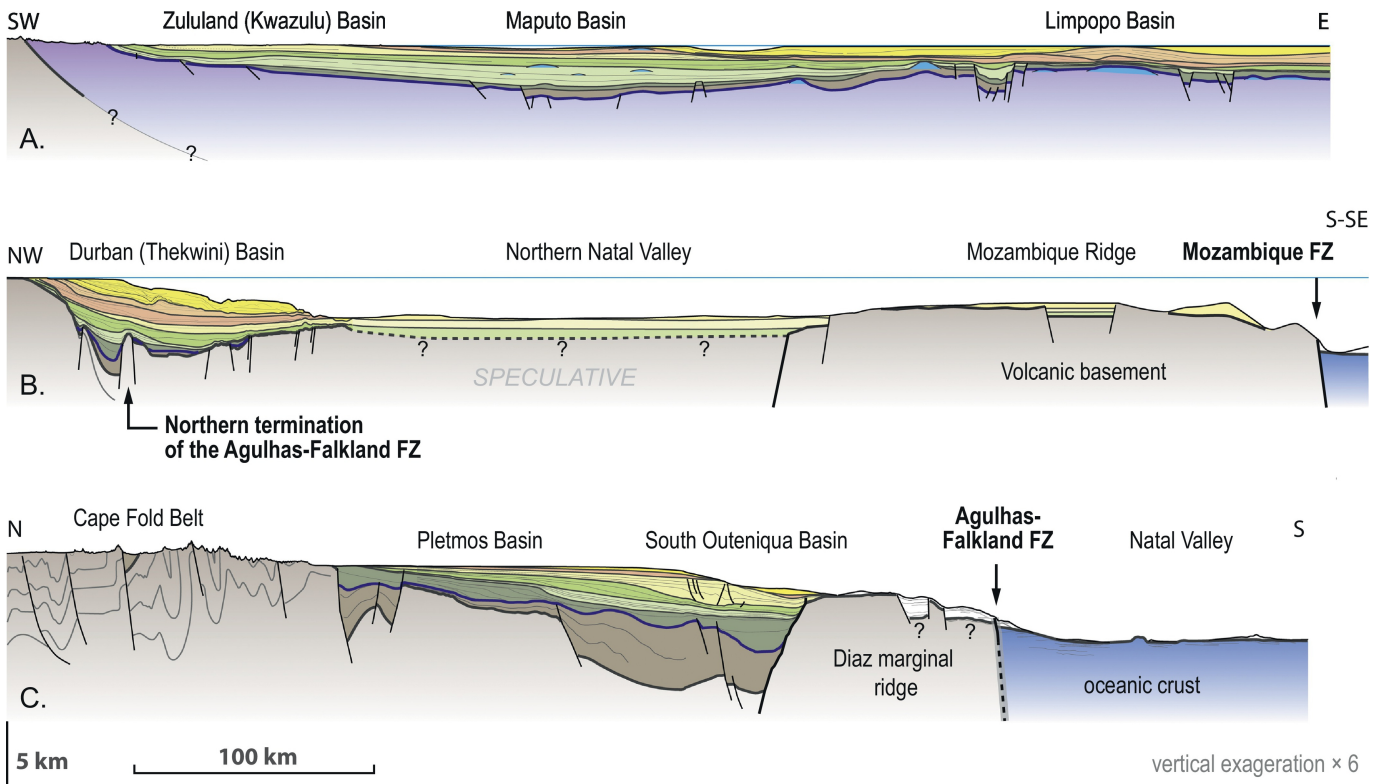


Figure 2

OFFSHORE

Post-Rift (Cretaceous-Cenozoic) isochore
c.i. 0.4 stwt



- ◊ Basement High
- ◊ Stratigraphic wells
- Stratigraphic outcrops

ONSHORE

Sediments

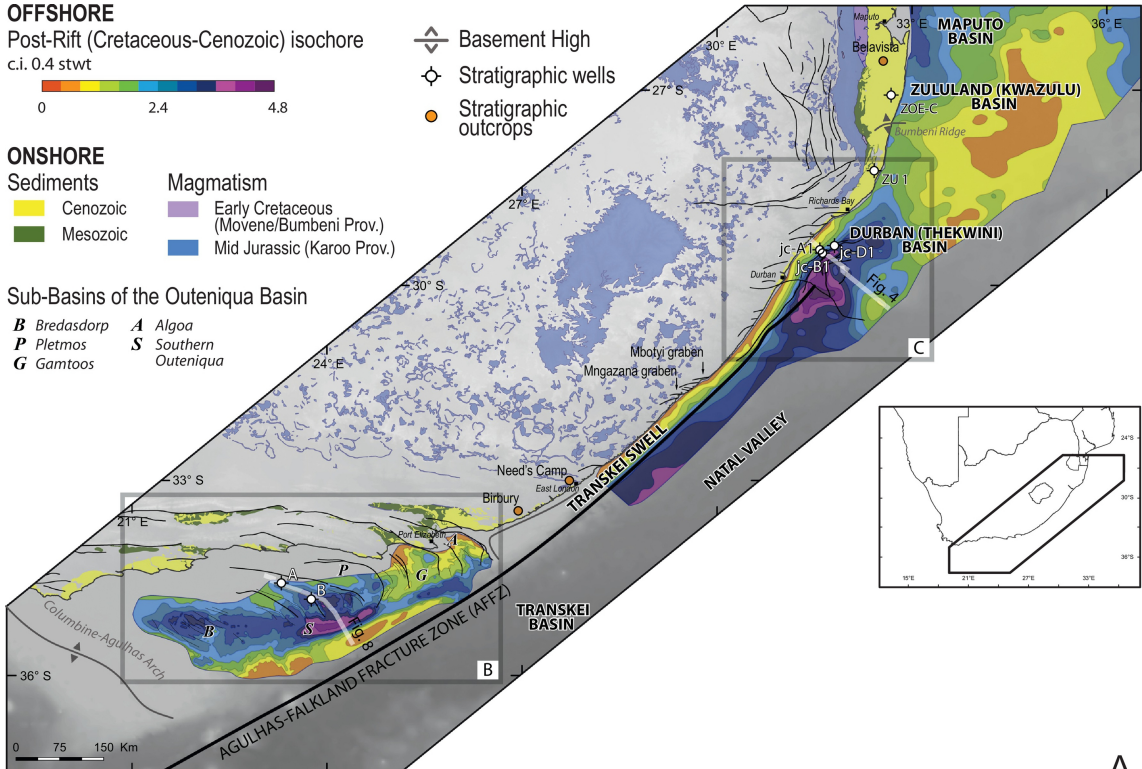
- Cenozoic
- Mesozoic

Magmatism

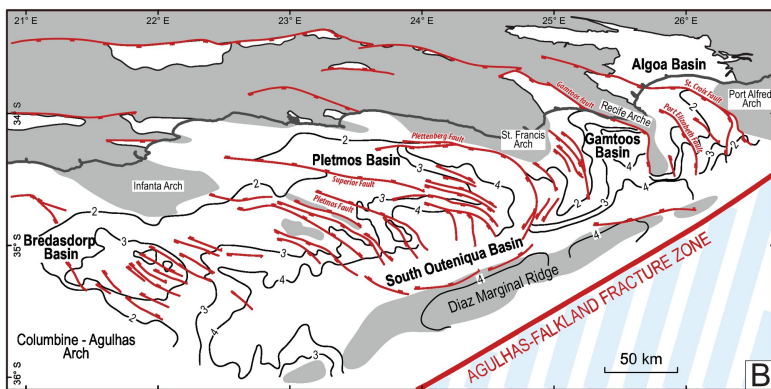
- Early Cretaceous (Movene/Bumbeni Prov.)
- Mid Jurassic (Karoo Prov.)

Sub-Basins of the Outeniqua Basin

- B** Bredasdorp
- P** Pletmos
- G** Gamtoos
- A** Algoa
- S** Southern Outeniqua



A.



- Shallow Basement
- Stretched Continental Crust
- Oceanic Crust
- 2- Basement isochrone (sTWT)
- Normal faults

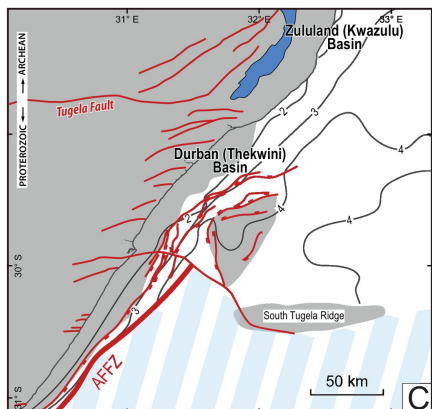


Figure 3

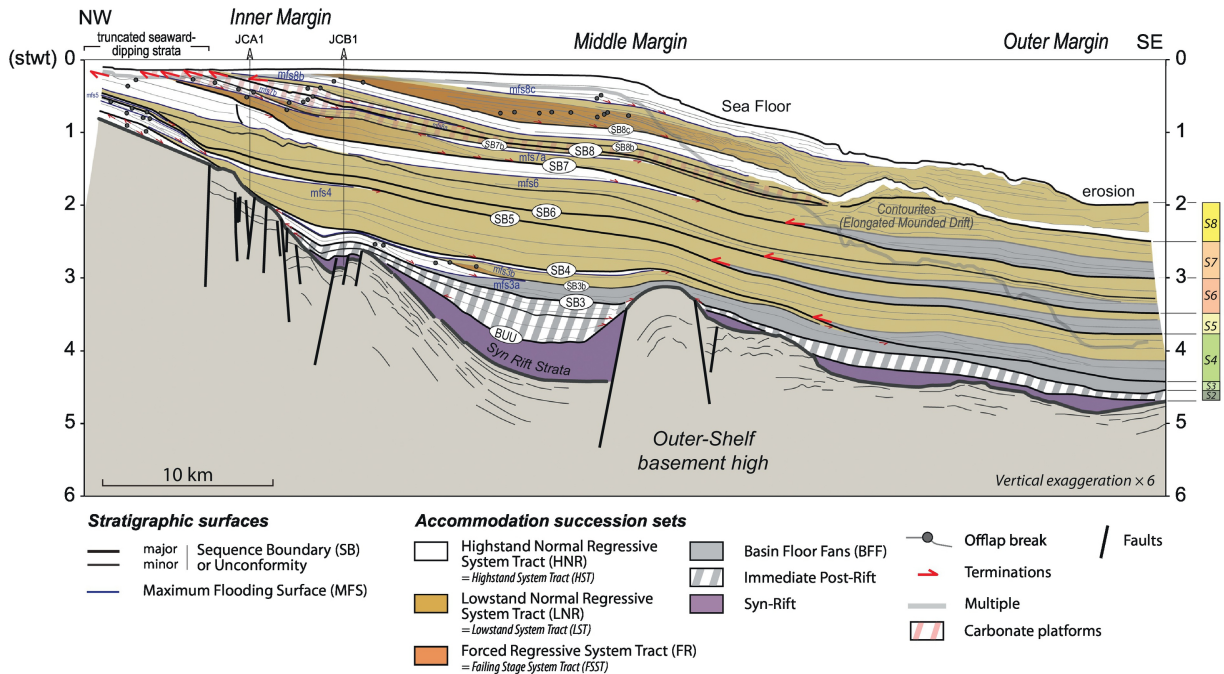
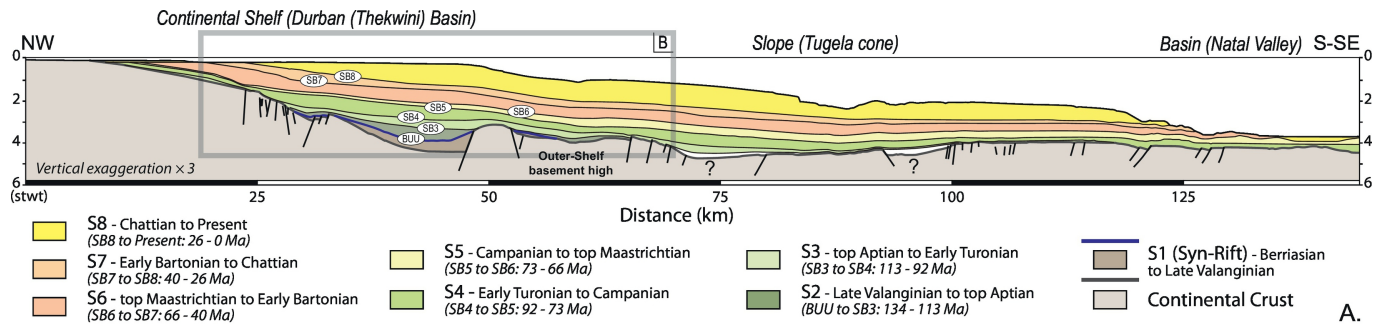


Figure 4

JC-A1 well

JC-B1 well

BIOSTRATIGRAPHY

SEQUENCE BOUNDARY CORRELATIONS

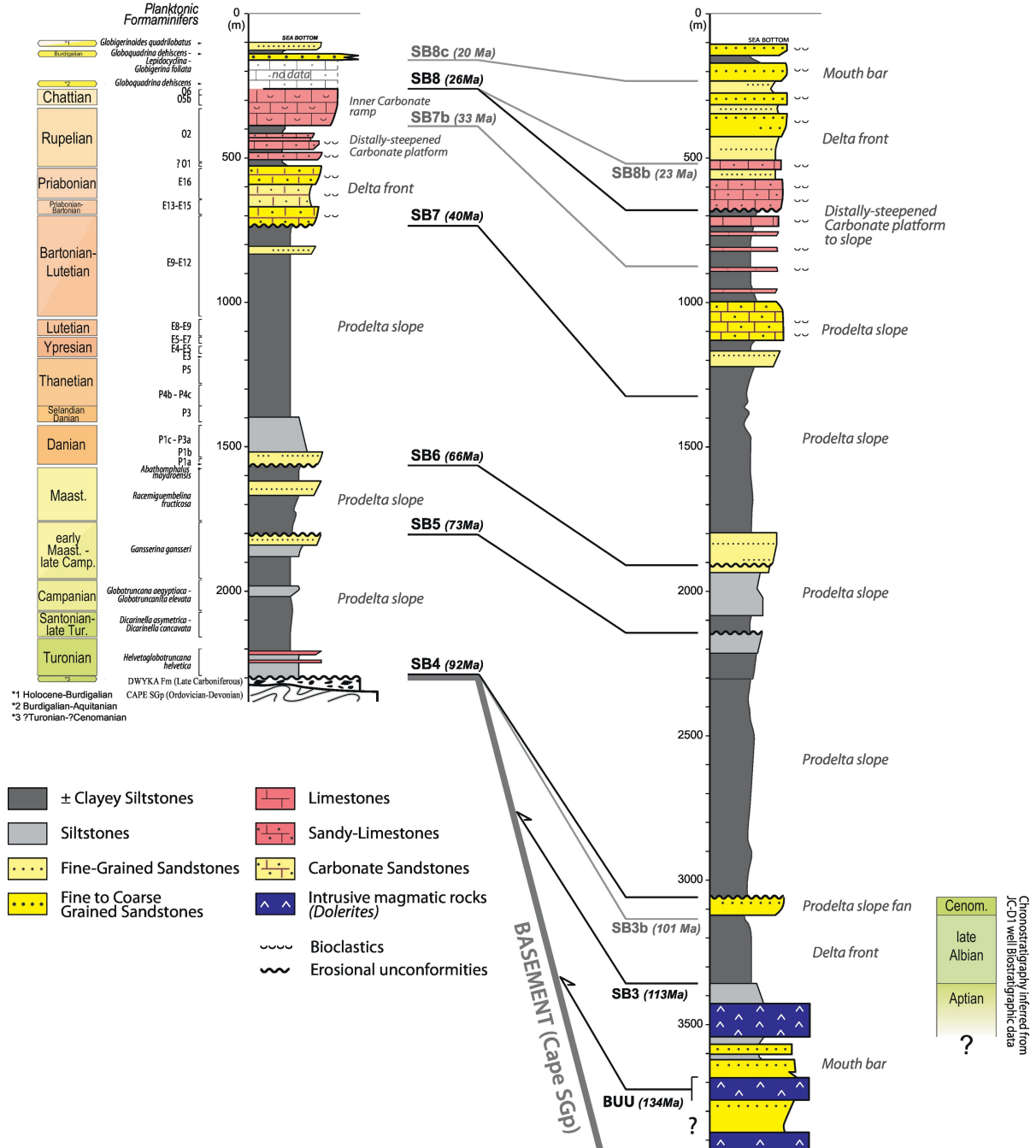


Figure 5

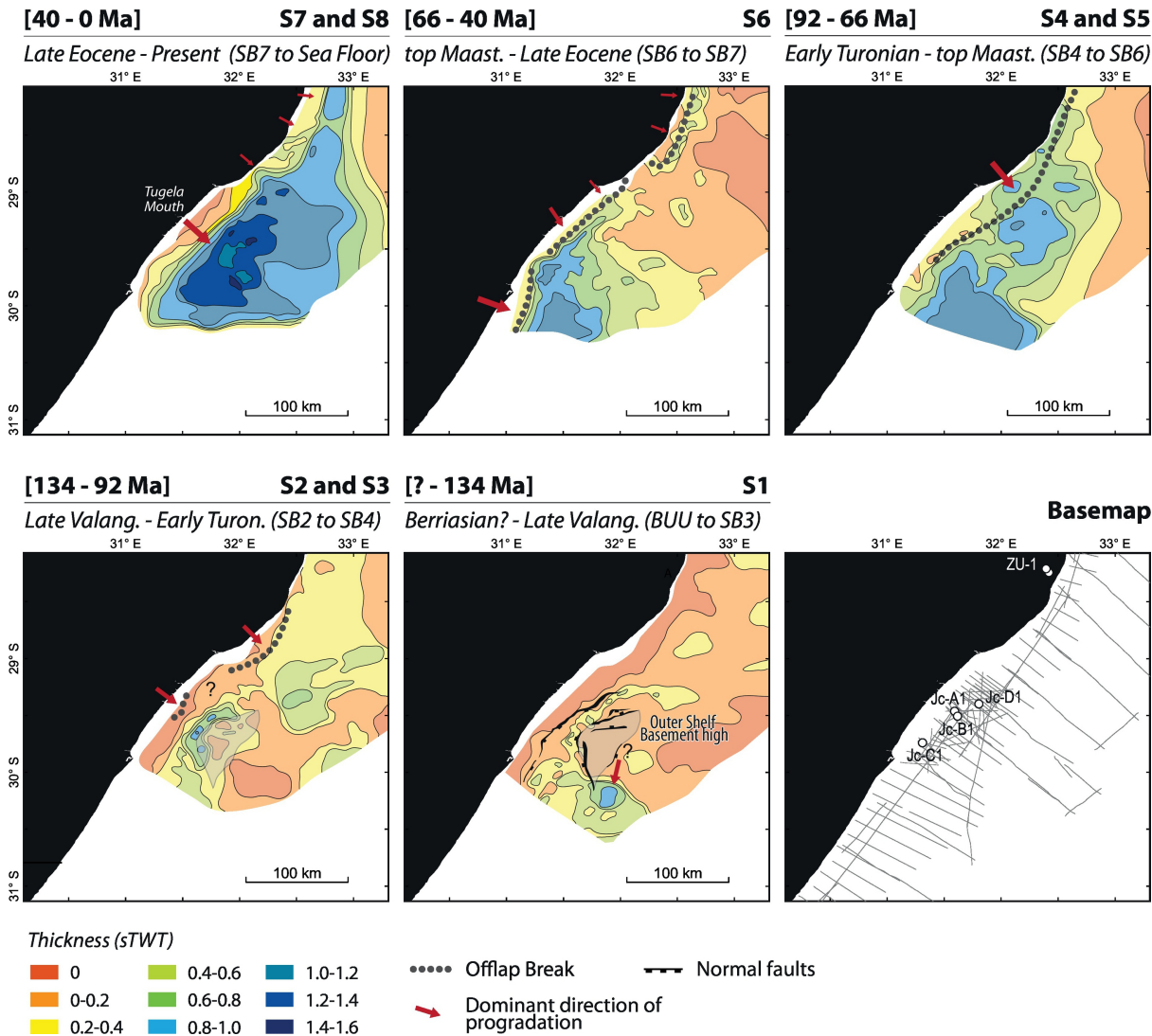


Figure 6

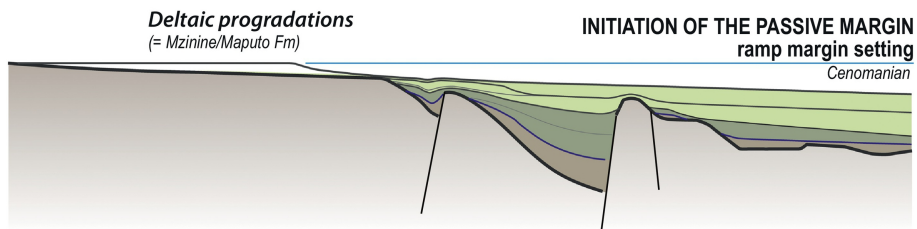
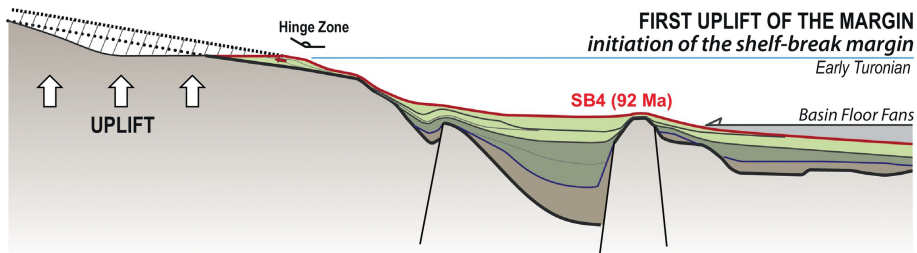
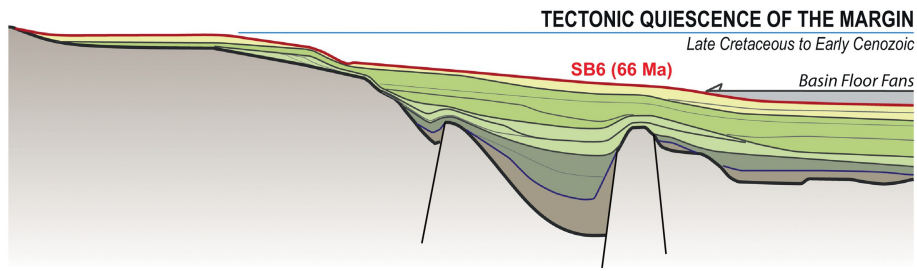
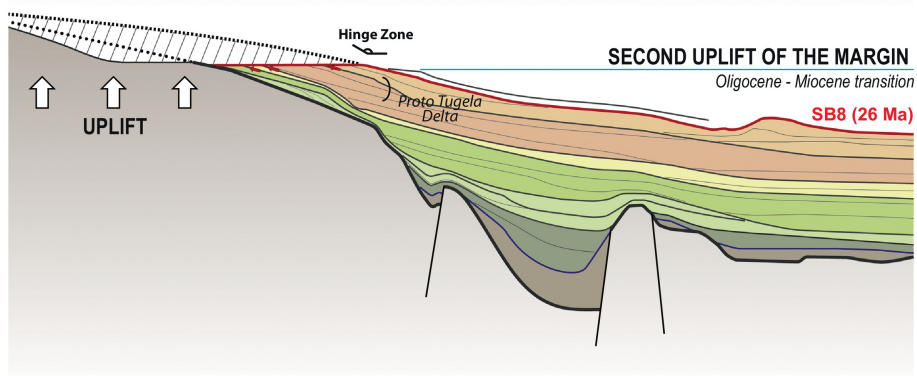
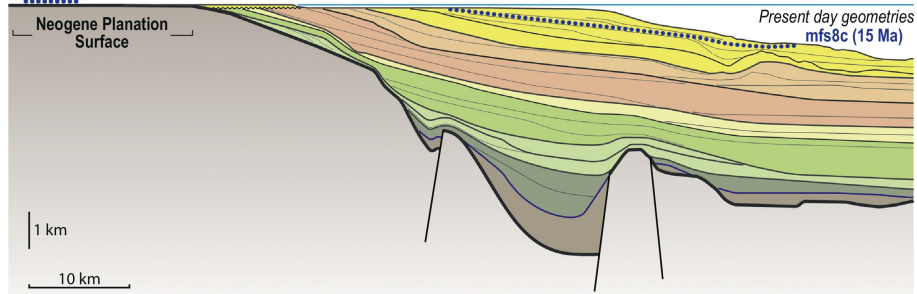
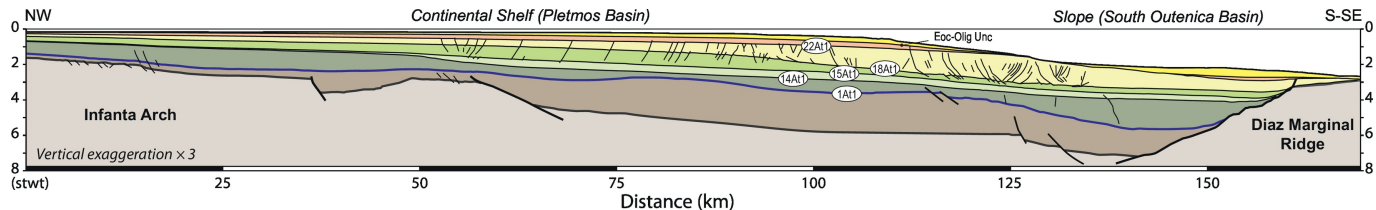
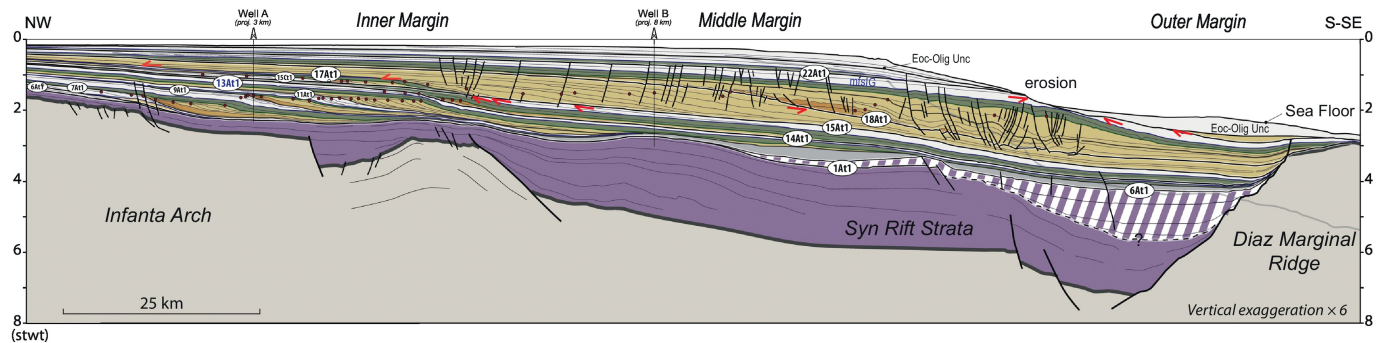


Figure 7



- | | | |
|--|---|--|
| S7 - top Oligocene to Present (23 - 0 Ma) | S4 - Early Turonian to Campanian (15At1 to 18At1: 92 - 76 Ma) | S1 (Syn-Rift) - Upper Jurassic to Late Valanginian |
| S6 - top Maastrichtian to top Oligocene (66 - 23 Ma) | S3 - top Aptian to Early Turonian (14At1 to 15At1: 113 - 92 Ma) | Continental Crust |
| S5 - Campanian to top Maastrichtian (18At1 to 22At1: 76 - 66 Ma) | S2 - Late Valanginian to top Aptian (1At1 to 14At1: 133 - 113 Ma) | |

A.



Stratigraphic surfaces

- Flooding Surface (MRS=TS=FS)
- major Sequence Boundary (SB) or Unconformity
- Maximum Flooding Surface (MFS)

Accommodation succession sets

- Highstand Normal Regressive System Tract (HNR) = Highstand System Tract (HST)
- Lowstand Normal Regressive System Tract (LNR) = Lowstand System Tract (LST)
- Forced Regressive System Tract (FR) = Failing Stage System Tract (FSST)

- Basin Floor Fans (BFF)
- Late syn-rift
- Syn-Rift

- Offlap Break
- Truncations
- Onlaps
- Multiple
- Faults

B.

Figure 8

Well A

Well B

AGES

SEISMIC MARKERS CORRELATIONS

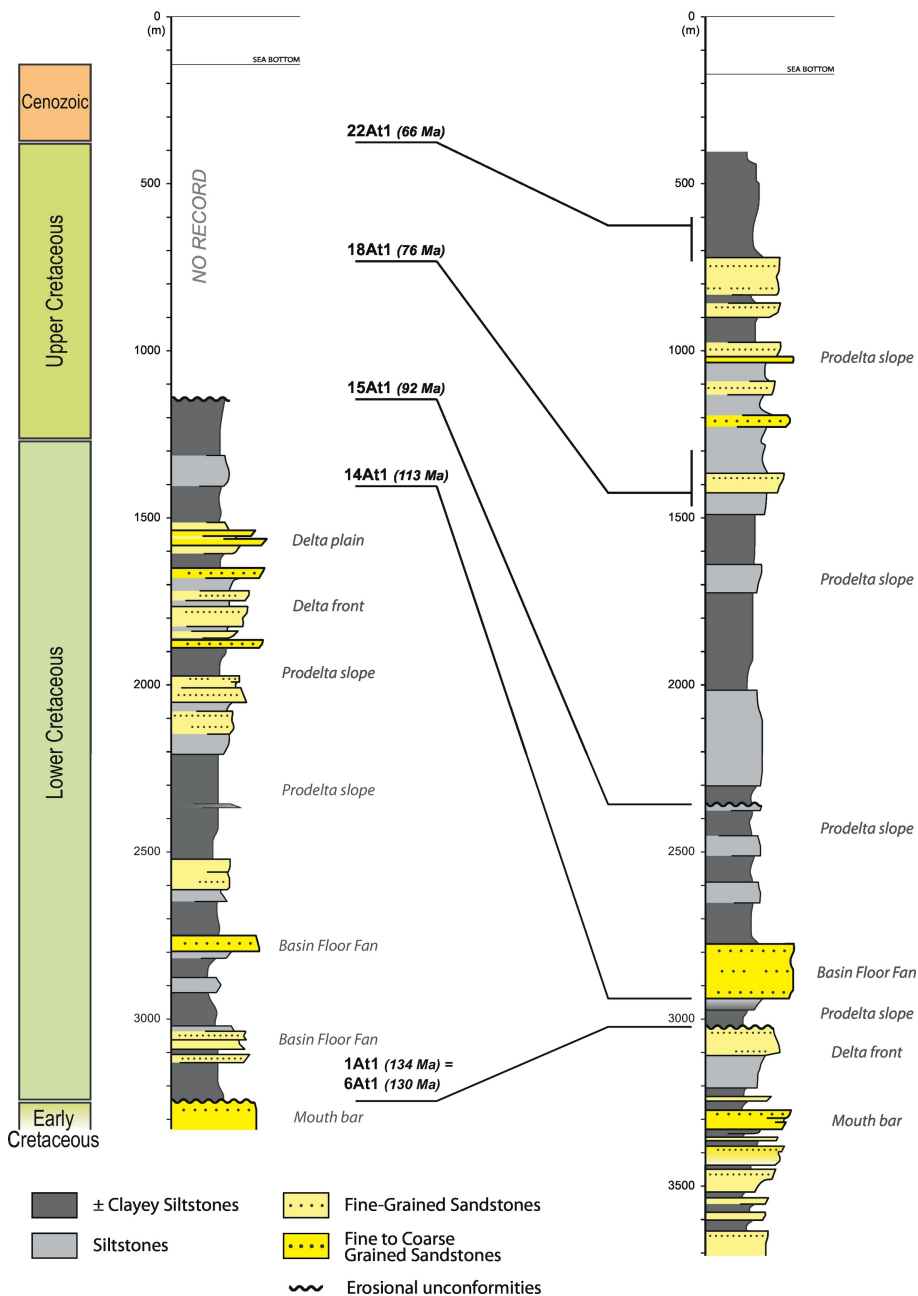


Figure 9

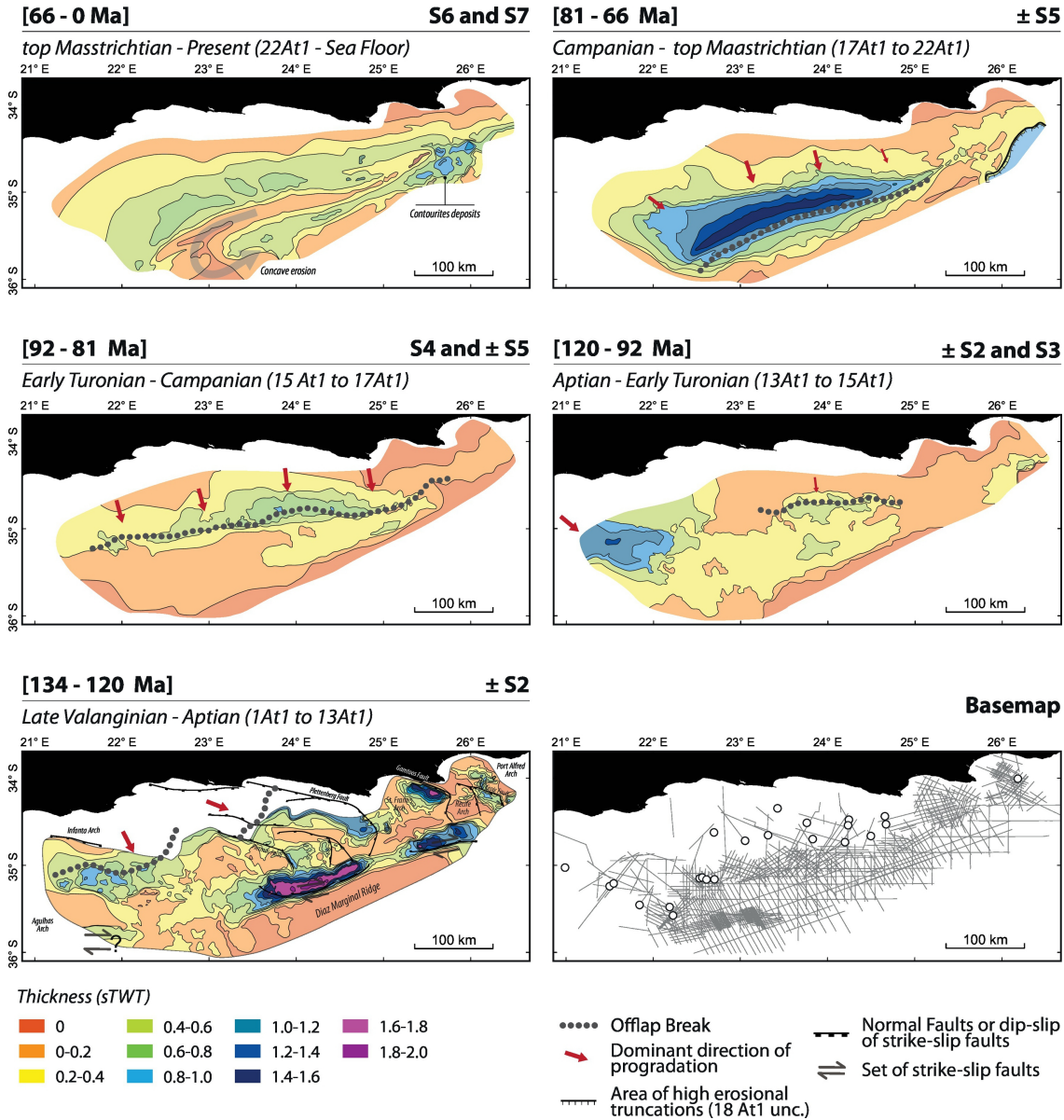


Figure 10

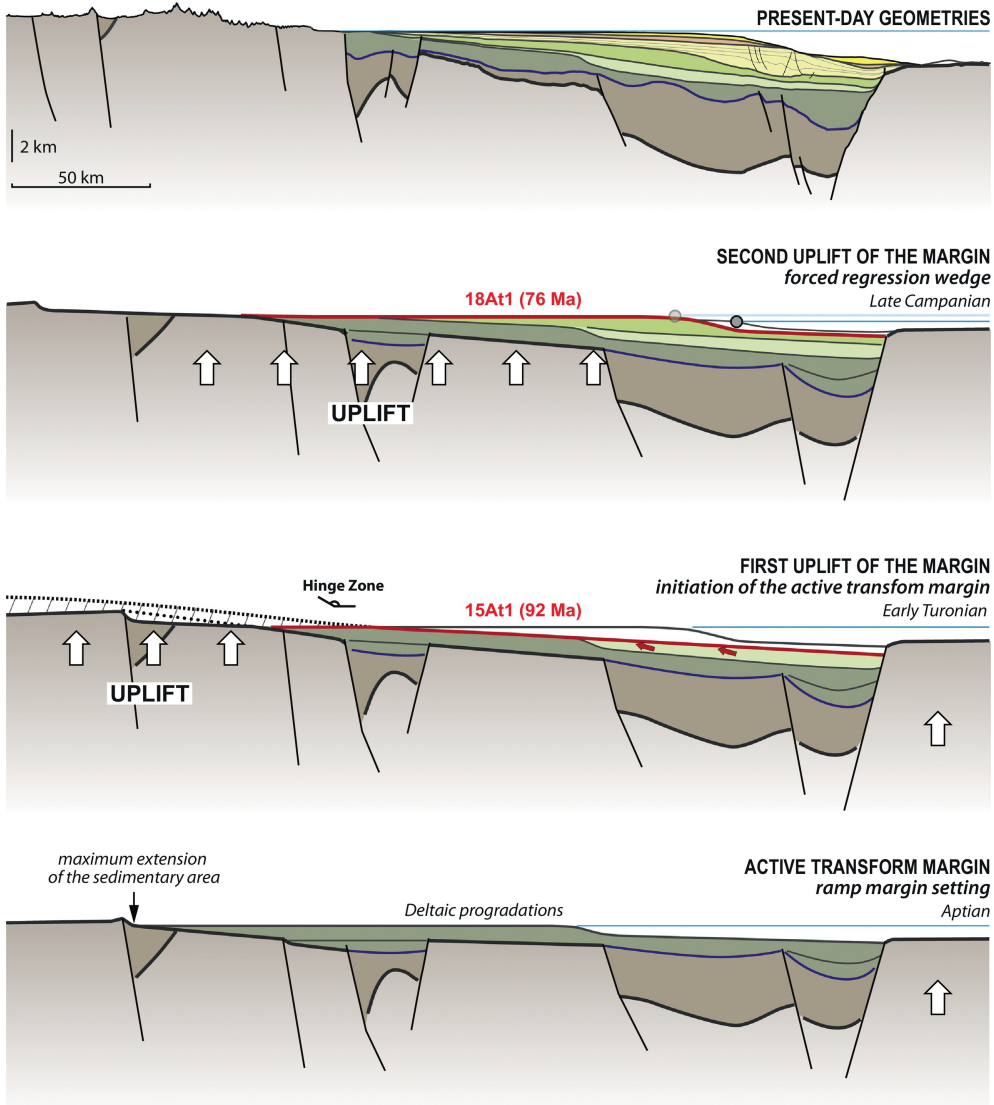
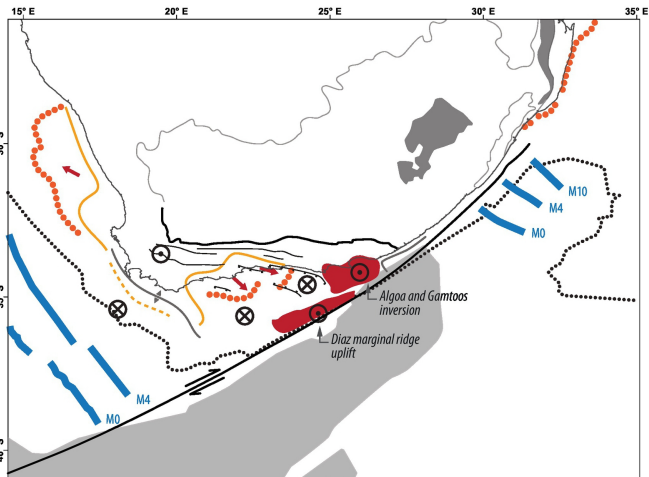


Figure 11

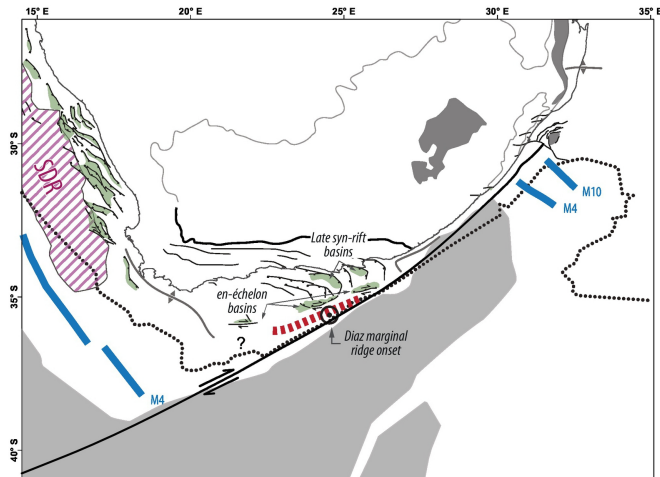
LATE HAUTERIVIAN - LATE APTIAN (131 - 113 Ma)

Active transform motion of the Agulhas - Falkland Fracture Zone



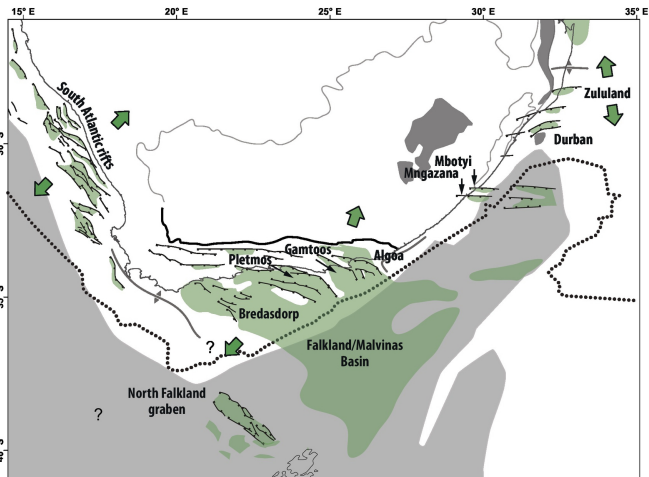
LATE VALANGINIAN - LATE HAUTERIVIAN (134 - 131 Ma)

Agulhas - Falkland/Malvinas transform margin initiation



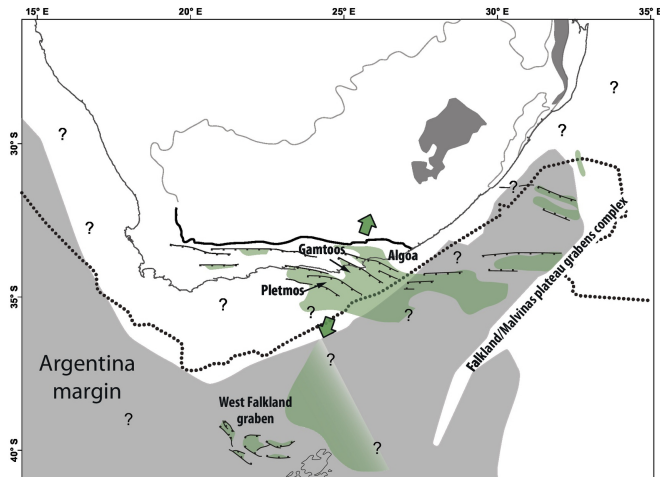
UPPER JURASSIC - LATE VALANGINIAN (150 - 134 Ma)

Gondwana break-up, third continental rifting phase



MIDDLE - UPPER JURASSIC (170 - 150 Ma)

Gondwana break-up, second continental rifting phase



Red dashed line: Zones of erosion

Yellow dashed line: Shoreline

Green shaded area: Rift infill (syn-rift marine and continental sediments)

Pink hatched area: Seaward-dipping reflectors (syn-rift volcanics)

Red dotted line: Delta shelf edge

Blue line: Magnetic anomalies

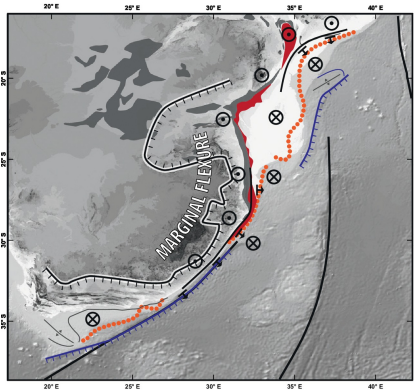
Circle with dot: Uplift

Circle with cross: Subsidence

Figure 12

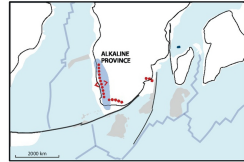
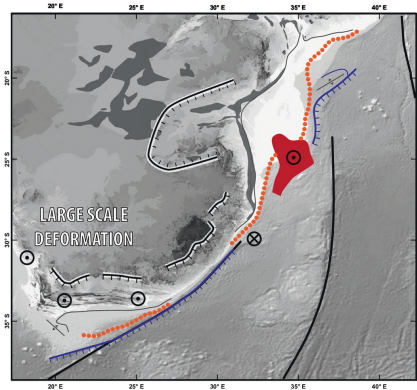
OLIGOCENE - EARLY MIOCENE DEFORMATIONS (34-15 Ma)

Second South-African Plateau Uplift



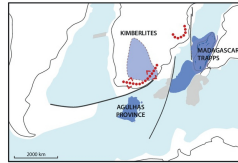
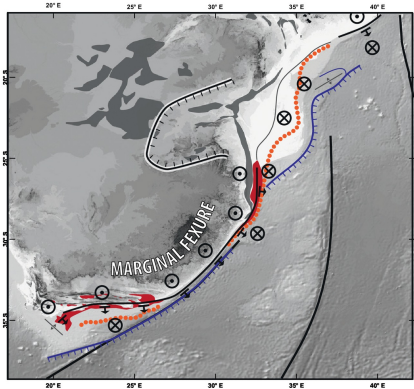
LATE CRETACEOUS DEFORMATIONS (81-70 Ma)

First South-African Plateau Uplift - late stage



EARLY LATE CRETACEOUS DEFORMATIONS (~92 Ma)

First South-African Plateau Uplift - early stage









-  Coastal Escarpment
-  Zones of erosion
-  Uplift
-  Subsidence
-  Shelf-Break
-  Delta shelf edge

Figure 13

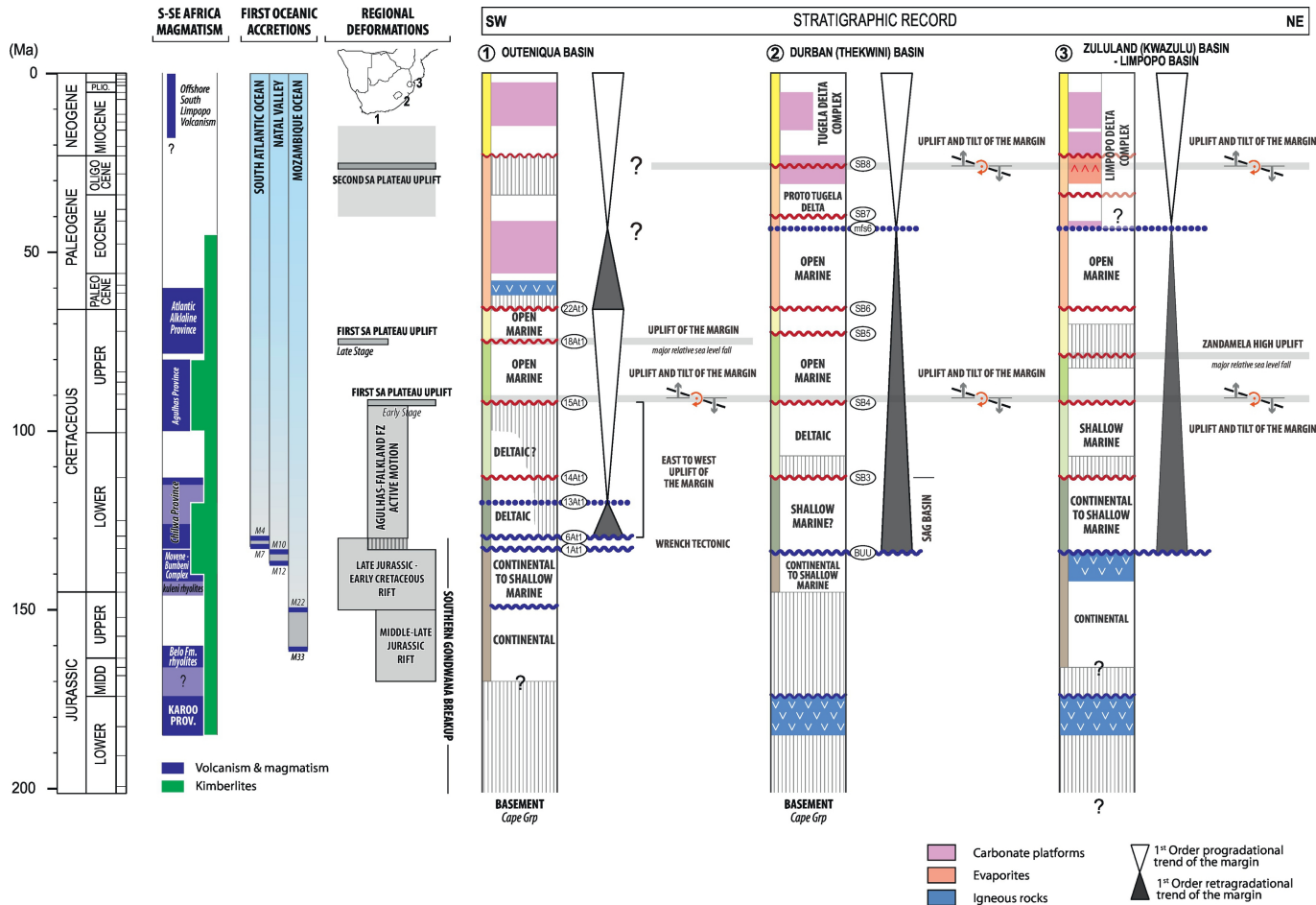


Figure 14